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THE ACADIAN TRIASSIC

SIDNEY POWERS
Troy, New York

PART II

Gerrish Mountain.—Gerrish Mountain consists of a basalt flow capping Triassic sandstones (Figs. 20, 22). The sandstones on the west side of the mountain are horizontal, those on the east side dip northward at angles of 15° or less.

Opposite Moose Island is a cross-section of the basalt flow (Fig. 22), showing sandstone faulted against agglomerate on the west, with the agglomerate overlain by sandstone in one exposure; a columnar basalt dike east of the agglomerate, and probably separated from it by a fault; and red sandstone, east of the dike, overlain by a 3-foot bed of ash, above which is columnar basalt. The dike is 50 feet thick. It extends northward for some distance, but it is cut off on the south by a fault between the mainland and Moose Island. In either the dike, or the basalt flow, magnetite has been mined near Lower Economy.

The final exposure of Triassic basalt on the east, which may originally have been connected with that at Gerrish Mountain, is at Portapique Mountain, 12 miles distant, and north of Birch Hill (see Fig. 23).

Gerrish Mountain–Truro.—The broad lowland underlain by Triassic strata east of Gerrish Mountain is interrupted by a long

strip of Carboniferous conglomerates and shales forming a high, rugged topography from Lower to Upper Economy, as shown in Figs. 19 and 24, rising 250 feet, or more, from the lowland on the south. The relation of the Triassic to the older rocks, as seen along the shore, is a fault west of Carr Brook, and an unconformity near Lower Economy (Fig. 25).

At the fault, the Triassic consists of very calcareous sandstones, containing conglomerate lenses and cross-bedding. Calcite has been introduced into the sandstone, forming dark-red concretions.

The unconformity near Lower Economy shows a basal conglomerate composed of subangular pebbles 1 to 3 inches in length, resting on upturned and leveled Carboniferous red shale. The shale within a foot of the contact is weathered into a clay, but this

FIG. 22.—The shore section along Gerrish Mountain, as seen from the eastern end of Moose Island. The Triassic sandstone, at the left, is faulted against a mass of agglomerate overlain by a capping of sandstone. The agglomerate is probably faulted against the dike of diabase, which has fed the basalt flow which caps the cliffs on the right. At the base of this flow is a bed of green ash, and beneath this is normal Triassic sandstone, with some shale.

weathering is of recent date. The basal Triassic conglomerate pebbles consist of slate, schist, quartz, and igneous rocks from the Cobequid Mountains. Above the basal conglomerate, which is 25 feet thick, are interbedded sandstones and conglomerates as seen in Fig. 26. It is 1,200 feet stratigraphically between the basal unconformity and the Gerrish Mountain basalt flow.

From Economy to Truro, Minas Basin is fronted by a comparatively low land underlain by practically horizontal Triassic sandstones, above which, in places, are Pleistocene gravels.

The northern contact is of importance because there is a question whether there is a fault or an unconformity. The exposures are confined to the steam valleys, and are quite unsatisfactory. It may be briefly stated that the relation is probably a fault as far as Chiganois River, with another fault from there to North River,

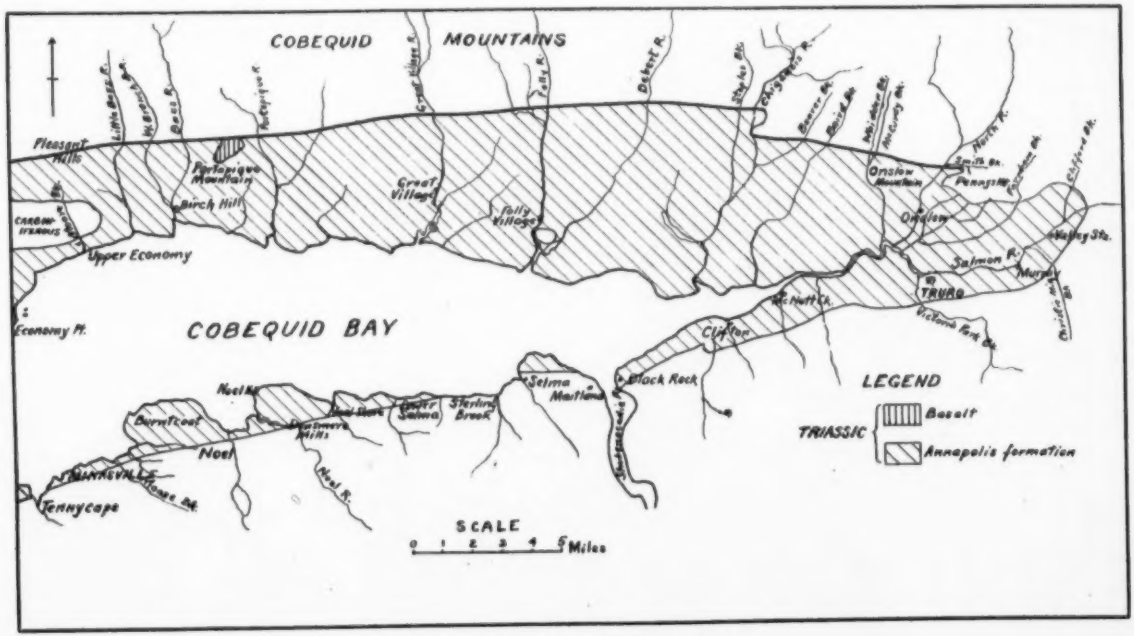

FIG. 23.—Map of the head of Minas Basin

as shown in Fig. 23. The evidences which favor this conclusion are the apparently dragged dips of the Triassic in several cases; and the lack of pebbles of adjoining older rocks in the Triassic conglomerates near by, and the actual fault seen in Harrington River on the west.[1]

At Birch Hill and at Folly Village, older rocks project through the Triassic strata according to Fletcher, who mapped both

FIG. 24.—The base of the Triassic at Minasville, showing the basal sandstones and conglomerates resting on horizontally trunkated Carboniferous shales. The character of the erosion surface indicates a peneplain, and the composition of the basal conglomerate indicates a lack of residual soil on this surface at the beginning of Triassic sedimentation.

localities with exaggeration of their size. There are reasons for questioning the existence of older rock at Birch Hill, but the copper prospect where these rocks are supposed to occur was not visited in the reconnoissance.

[1] Certain of the localities are described by H. Fletcher, *Geol. Surv. Canada, Ann. Rept.*, V (1892), 142–43 P.

At Folly Village, Mississippian fossiliferous limestone and gypsum, belonging to the Windsor group, appear on the northwest side of Debert River. The Triassic sandstones are very calcareous, and resemble the Windsor limestones, as both are red in color. It appears that the limestone is overlain conformably by the gypsum, and that these Mississippian strata are overlain disconformably by the Triassic red sandstones. With this view, the Mississippian is confined to a small area on the northwestern side of Debert River where fossils are readily found.[1]

Truro–Wolfville.—The end of the arm of Triassic in Minas Basin lies near Truro, and the relation of these to the older rocks

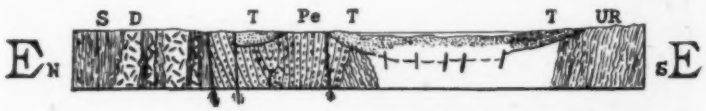

FIG. 25.—Section EE. Structure section through Minas Basin near Bass River, to show the unconformity of the Triassic on an island of Permian strata, Parrsboro formation, just north of the Basin. *SD*, Cobequid group; *UR*, Union-Riversdale series (Pennsylvanian); *Pe*, Permian. The major Cobequid fault is shown at the south of the Cobequid group. The closely folded syncline of the Parrsboro formation is in part overlain by Triassic sediments (*T*), which appear to be down-faulted on the north.

is an unconformity, as shown in Fig. 23. This unconformity is well exposed in Salmon River and in Victoria Park Brook (where there is also a fault). The underlying Carboniferous strata always show a beveled surface. This unconformity continues along the south shore of Minas Basin, and may be seen at Minasville, on the sides of Moose Brook (Fig. 24), at Tennycape, and at Walton. Over this area the Triassic sandstones show nothing unusual, except for calcitization north of Maitland.

West of Cheverie, the first exposure of Newark rocks is at Oak Island, north of Avonport (Fig. 3). On the east side of this island, quartz-pebble conglomerate and sandstone are exposed, overlain by stratified Pleistocene gravels, 6 feet thick, above which is Wisconsin till.

[1] This view differs from that of J. W. Dawson, *Acadian Geology*, 3d ed., 1878, p. 99.

North of Oak Island is Boot Island, which is separated from the mainland, called Long Island, by a narrow channel, formed within the last two centuries. North of this channel is a buried forest, exposed at low tide. North of Boot Island, and on the north side of Long Island, are exposures of red sandstones with occasional thin shales in which Dr. H. M. Ami reports the presence of *Estheria ovata*.[1]

FIG. 26.—Details of the Triassic sandstone and conglomerate 30 feet above the base of the Annapolis formation, one mile east of Lower Economy.

At Wolfville the basal Triassic unconformity is exposed in a small brook west of the buildings of Acadia College, resting horizontally on upturned and beveled slates of the Meguma series.[2] The basal unconformity is again exposed at Kentville, just below the mill on Black River.[3]

[1] Verbal communication.

[2] The writer is indebted to Professor Ernest Haycock of Wolfville for pointing out this locality.

[3] J. W. Dawson, *op. cit.*, p. 92; L. W. Bailey, "Geology of Southwestern Nova Scotia," *Geol. Surv. Canada, Ann. Rept.*, IX (1898), 128 M.

Wolfville–Scots Bay.—The Wolfville sandstone outcrops on the exposed points between Wolfville and Pereau River, the most continuous exposure being near Kingsport. The sandstone contains occasional conglomerate beds and red shales. The proportion of shale to sandstone gradually increases toward Blomidon. The shale in this locality is largely a red clay, with occasional green bands, persisting horizontally throughout the exposure. The general dip of the strata is 5°–10° northward. Small faults are numerous.

About midway between Kingsport and Pereau River, Haycock found fragments of well-consolidated fossiliferous red shale in till, overlying the Triassic.[1] The fossils are *Estheria ovata*. Dipping under North Mountain are poorly consolidated Blomidon shales, with the characteristic thin green beds at distances of 10–20 feet.

The Blomidon shale continues around the hook of North Mountain beyond Cape Blomidon, but not as far as Amethyst Cove. At the latter locality, basalt cliffs, partly columnar, rise abruptly from the sea to a height of 300 to 400 feet. These cliffs are kept vertical by frost action on the vertical joint planes parallel to the shore.

Two basalt flows are visible at Amethyst Cove, dipping gently northward, with undulating folds. The collecting place for amethysts is in a greatly veined area about 100 feet below the top of the lower flow.

Scots Bay–Bennetts Bay.—In the region around Scots Bay there are two points of especial interest: first, the origin of the curve in North Mountain at this point, and, secondly, the presence of the Scots Bay formation overlying the North Mountain basalt along the southeast side of the Bay. Furthermore, the structural evidence furnishes a clue to the former thickness of the younger formation.

The curve in North Mountain is formed in a syncline pitching down to the west, and in the nose of this syncline Scots Bay has been eroded (see Figs. 27, 28). The basalt flows of North Mountain dip toward the Bay on all sides at angles of 3°–5°. The topographic

[1] E. Haycock, "Fossils in the Boulder-Clay of Kings County, Nova Scotia," *Trans. N.S. Inst. Sci.*, X (1901), 376–78.

slope follows the dip slope closely, beveling it slightly. At the water's edge there is no sea-cliff, but merely a sheet of basalt (where exposed) sloping upward from the shore. Farther southwest along North Mountain there are low sea-cliffs, but the general dip-slope persists to the end of Brier Island. The crest of North

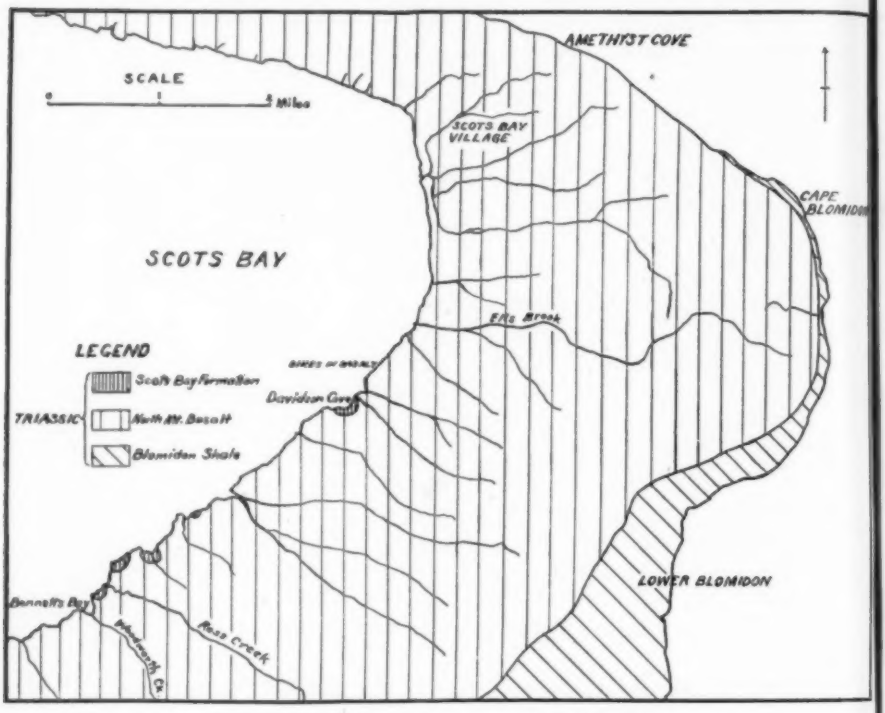

FIG. 27.—Map of the Scots Bay–Cape Blomidon region

Mountain is a mile or two southeast of the Bay of Fundy shore. It is a uniformly rolling surface which bevels the tilted basalt flows and forms a remnant of the Summit peneplain.

Around Scots Bay the top of the uppermost basalt flow is marked by a green amygdaloidal layer in which the amygdules are one-half to three-quarters of an inch long. This amygdaloid

outcrops on the shore from Bennetts Cove to Ells Brook (Fig. 27). From this point to the north shore of the Bay there are no outcrops. Along the north shore toward Cape Sharp, the amygdaloid has been eroded, exposing the solid basalt beneath. In places this basalt is columnar, and in other places it contains balls, a foot or more in length, composed of dense basalt surrounded by amygdaloidal rims. The balls are not sufficiently abundant for the basalt to be called a pillow lava.

The younger, sedimentary formation, named for its occurrence on the south side of Scots Bay, the Scots Bay formation, has a thickness of 25 feet. It was discovered by Ells in 1876, and described by him[1] and later by Haycock.[2] It consists of white, very calcareous sandstone, quite distinct in color from any other Triassic sandstone of Acadia, with some interbedded shale and normal sandstone.

The Scots Bay formation outcrops in five small synclines between Scots Bay and Bennetts Bay (formerly called Woodworth Bay). These small remnants rest conformably on the basalt, contrary to the opinion of Haycock, and are preserved in small synclines in the basalt. They extend southeast only a few hundred feet, as the topographic slope of the

[1] R. W. Ells, "Notes on Recent Sedimentary Formation on the Bay of Fundy Coast," *Trans. N.S. Inst. Sci.*, VIII (1894), 416–19.

[2] Ernest Haycock, "Records of Post-Triassic Changes in Kings County, Nova Scotia," *ibid.*, X (1900), 287–302.

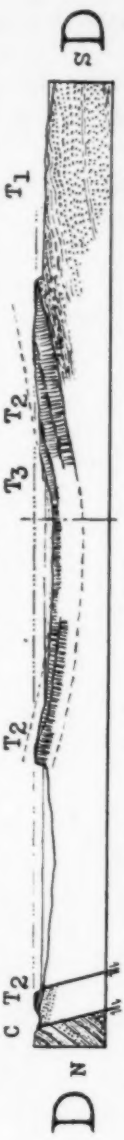

FIG. 28.—Section DD. Geologic cross-section through Cape Sharp on the north; Cape Split, Scots Bay, and Cape Blomidon, on the south. *C*, Carboniferous; T_3, Triassic, Annapolis formation; T_2, North Mountain basalt; T_1, Scots Bay formation. The vertical scale from Cape Split southward is exaggerated to show the overlying Scots Bay formation which is only 25 feet thick. The level of the Summit peneplain is shown by the line with three dots and a dash, the top of the flows of North Mountain by the finely dashed line. The area between these two lines must have been filled with the Scots Bay formation at the time the Summit peneplain was developed.

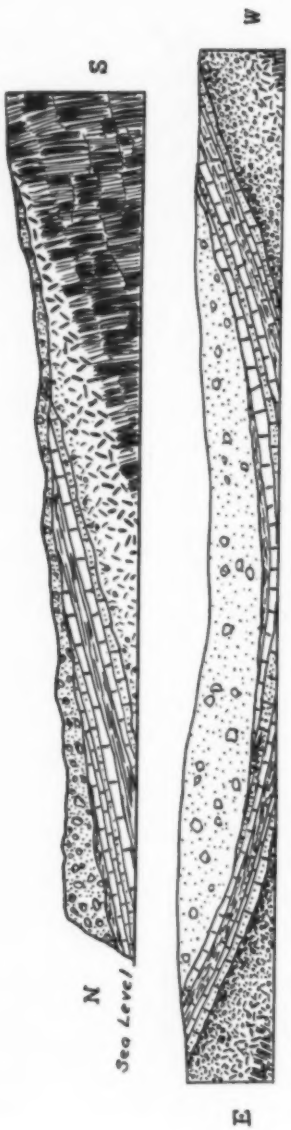

FIG. 29.—Structure sections of a syncline of the Scots Bay formation, lying conformably over the top of the North Mountain basalt. Above the Scots Bay formation is till, filling the pre-Wisconsin valley. The top of the upper flow is very amygdaloidal. The basal beds of sandstone have been replaced by chert.

mountain cuts across them and the flows at a low angle. These synclines are shown in Fig. 27.

The Scots Bay formation consists of calcareous white or gray sandstone, frequently replaced by chert, and greenish sandstone or shale. The exposures are nowhere over 15 feet in thickness (Fig. 29) and are remnants of a formation which once filled the syncline of Scots Bay up to the level of the Summit peneplain, as shown in Fig. 28. The white sandstone, or chert, rests directly on the amygdaloid at the top of the basalt flows, and veins of chert run downward from the sandstone into the amygdaloid. The beds in any one syncline do not match exactly with those in any other, but this condition is to be expected in basal beds of which only a few feet are shown, on the irregular top of a lava flow.

Fossils have been found in the calcareous sandstones by Haycock. They consist of faint green markings, probably plant remains; worm burrows; fish scales, bones, coprolites, and other

fragments of fish. The coprolites are from 1 to 1½ inches in length, an inch wide, and half an inch thick. In 1913, Haycock found a portion of the head of a fish which has been identified by Mr. L.M. Lambe as the Triassic genus, *Semionotus fultus* (Agassiz).

Digby Gut.—Between Kentville and Digby Gut there are few outcrops of the Annapolis formation and no cross-sections of the North Mountain basalt. The best cross-section of the latter is at

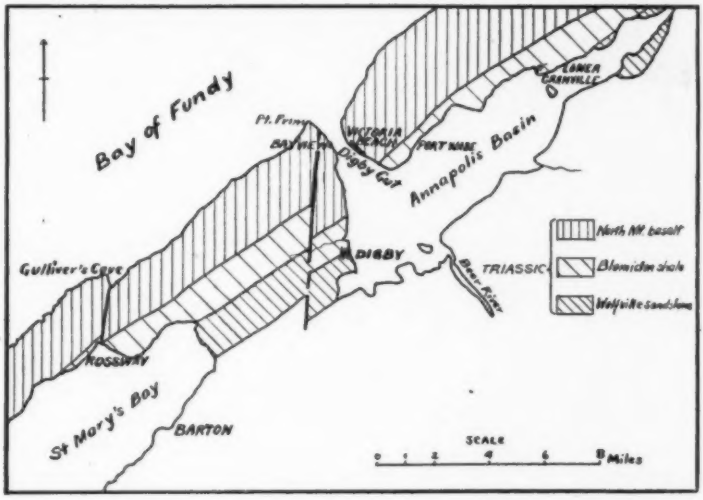

FIG. 30.—Map of the Digby-Rossway region

Victoria Beach, on the east side of Digby Gut (Fig. 30). The exposures on the west side of the Gut are disturbed by faulting.

The section of North Mountain basalt, except the lower flow, which appears at the side of Digby Gut, commences on the Bay of Fundy shore, where 7 flows may be seen (see Fig. 31), each having a thickness ranging from 2 to 45 feet. All the flows dip toward the Bay of Fundy at a low angle, as seen in Fig. 32, but this uniform slope is in places interrupted by minor folds. The presence of a low syncline and the lack of exposures make the exact thickness of the lower flow uncertain, but it is probably about 600 feet.

Faults occur at right angles to the axis of North Mountain, at Digby Gut and at Bay View (Fig. 33). The physiographic evidence of these faults is seen in the offset of the North Mountain basalt, and in the valleys along the fault-lines.

On the shore of Annapolis Basin near Port Wade, and at Digby, there are exposures of slightly cemented sand, containing blocks

FIG. 31.—Five lava flows of North Mountain at Digby Gut, as seen north of Victoria Beach, on the northeast side of the Gut. The upper flow is the third from the top of the series as exposed.

of basalt, which were considered by Bailey to be of Triassic age, and to underlie the basalt flows of North Mountain.[1]

These beds are of post-Wisconsin age, because of (1) the lack of consolidation, except very locally; (2) the yellow color, like ordinary stream gravels (unlike any Triassic deposit except the

[1] L. W. Bailey, "Triassic (?) Rocks of Digby Basin," *Trans. N.S. Inst. Sci.*, IX (1898), 356–60; also "Geology of Southwestern Nova Scotia," *Geol. Surv. Canada*, IX (1898), 126 M.

FIG. 32.—The section through North Mountain on the northeast side of Digby Gut. The village of Victoria Beach is seen on the opposite shore. North Mountain consists of basalt flows dipping to the north at approximately the angle seen in the dip-slope at the extreme left. The base of the flows is just beneath the landslide on the mountain near the center of the picture (as shown by dashes). Below and to the right of this landslide is red shale which has been weathered back to produce the escarpment on the south side of North Mountain.

Quaco conglomerate); (3) the lithological character and the lack of green laminae; (4) the delta character of the deposit at Digby; (5) the basalt fragments decreasing in number with increasing distance from the talus of North Mountain basalt; (6) the horizontality of the stratification (the red and green Triassic shales near Victoria Beach dip northward at an angle of 3°).

Rossway–Brier Island.—At Rossway on St. Mary's Bay (Fig. 30), there are excellent exposures of red shale, with occasional green beds and red sandstone beds, comprising the Blomidon shale. Rossway and Gulliver's Cove are connected by a valley which marks a north-south fault similar to that at Bay View. The displacement of the fault may be seen at Gulliver's Cove, in columnar basalt. No thin flows are shown here, or along Digby Neck to Brier Island.

The Blomidon shales at Rossway have a thickness of about 500 feet. They dip northward at angles up to 10°. No other shales in the Acadian Triassic are as well consolidated. Ripple marks, current undulations, cross-bedding, and rarely mud cracks are seen in the shales.

On the shore of Digby Neck, west of Rossway, the shales are exposed for a mile, dipping under the basalt. Half a mile west of the first exposure are Pleistocene clays containing black lignite and basalt fragments. This clay may have been deposited in the post-Wisconsin submergence.

Digby Neck, Long and Brier islands show in common a depression in the center of the ridge, parallel to the strike of the lava flows. This depression marks the amygdaloid at the top of the lower flow, as portions of but two flows are shown above the sea. No sedimentary rocks are shown on the St. Mary's Bay side west of the exposure near Rossway.[1]

Cross-faults are shown between Digby Neck and Long Island, and between Long Island and Brier Island. Another fault probably occurs between Brier Island and the submerged ledge on the

[1] A. Gesner described red sandstone exposed off Brier Island at low tide, but he probably mistook either the red seaweed or a hematite stain over basalt for sandstone (*Remarks on the Geology and Mineralogy of Nova Scotia*, Halifax, 1836).

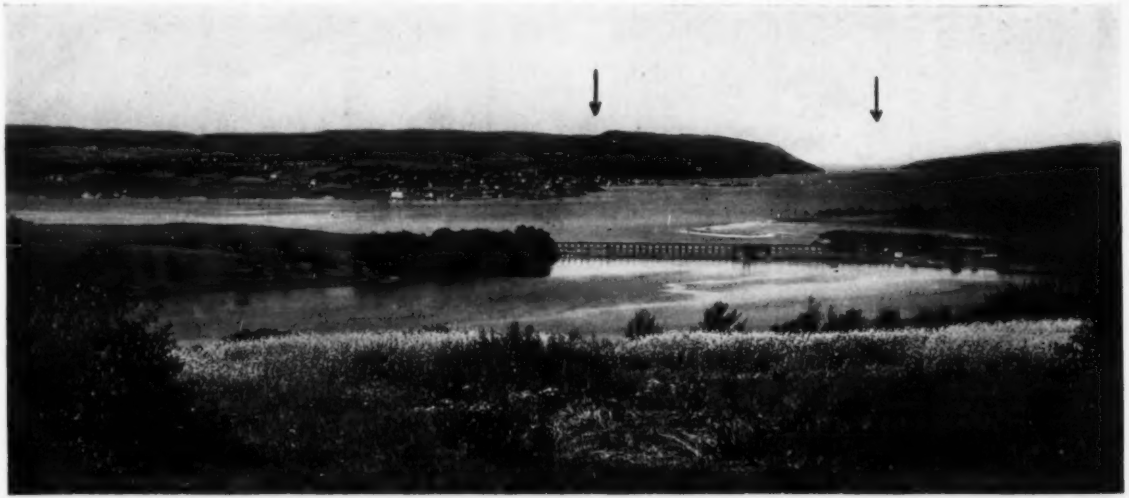

FIG. 33.—Digby Gut and Digby from the foot of South Mountain. The top of North Mountain, in the distance, marks the Summit peneplain. The arrow on the left points to the notch which marks the Bay View fault. The arrow on the right points to Digby Gut, which was probably formed on a similar fault-line. The block between the arrows stands relatively nearer Digby than the remainder of North Mountain on the left, because the downthrow of the cross-faults is on the southwest.

west. Beyond this ledge there is no evidence of the existence of any Triassic rocks.

AGE

The age of the Newark group must be determined by a comparison of its fauna and flora with that of Europe where the Triassic system is well developed. As forms common to both countries are not abundant, there are slight differences of opinion as to the exact correlation. Table I gives a correlation scheme which is modified from one given by Eastman.[1]

Trias	Great Britain	Germany	Eastern U.S.
	Rhaetic	Rhaetic	
Upper	Keuper marl Keuper marl (Upper Keuper sandstone)	Upper, Middle, Lower } Keuper	Newark group
Middle	Lower Keuper sandstone	Lettenkohle	
		Upper, Middle, Lower } Muschelkalk	
Lower	Upper variegated sandstone Pebble beds Bunter sandstone	Buntersandstein	

On the evidence of the fish fauna Eastman[2] concludes that the Newark is to be correlated with middle and upper divisions of the Alpine Trias (the Upper Muschelkalk–Middle Keuper of the German section). The plants indicate a similar age, and several forms have been cited as the equivalent of the Lettenkohle forms of Germany.

In the Acadian area the fossils which have been found are: plant remains at Split Rock (Gardner's Creek), Quaco, and Martin Head, New Brunswick; fish remains in the Scots Bay formation at Scots Bay; and impressions of the shells of bivalved crustaceans in drift material from the Blomidon shale, found near Kingsport.

[1] C. R. Eastman, "Triassic Fishes of Connecticut," *Conn. Geol. and Nat. Hist. Surv., Bull. 18*, 1911, p. 26.

[2] *Ibid.*, p. 29.

The plant remains were described by Dawson[1] from poorly preserved material as *Dadoxylon Edvardianum*. They consist of silicified plant-stems and of lignite, showing pith-casts. At Gardner's Creek and at Vaughan Creek (Quaco), the material is largely silicified, and appears to have been transported some distance. At Martin Head, lignite is exposed in several horizons and is quite abundant. Silicification has not replaced the plant tissues to such an extent as in the other localities.

The Martin Head locality is the only one where material is available for study. Miss Holden has recently examined the lignite, and found two species of plants.[2] The form which was assigned by Dawson to the genus *Dadoxylon* has been found to be identical with *Voltzia coburgensis* Schaur., from the Lettenkohle and Lower Keuper of Germany. The other form is *Equisetum rogersii*, Schimper, which has been described by Fontaine from the Virginia Triassic.[3]

The correlation of these forms is also considered by Miss Holden. The *Voltzia* is apparently the same as the form described by Newberry as *Palissya* from the New Jersey area,[4] and as the form *Cheirolepis* from New Jersey and Virginia. The *Equisetum rogersii* is probably identical with *E. columnaris*, described by Bronn, from the Lettenkohle.

Fragmentary fish remains have been found by Haycock[5] at Scots Bay in the calcareous sandstones overlying the basalt. Recently further collections have been made by Professor Haycock and the material has been identified by Mr. L. M. Lambe, of the Geological Survey of Canada, as probably *Semionotus fultus* (Agassiz), a form common to the other Newark areas.

[1] J. W. Dawson, *Acadian Geology*, 3d ed., 1878, p. 108; also, *Notes and Addenda*, p. 99.

[2] Ruth Holden, "Fossil Plants from Eastern Canada," *Annals of Botany*, XXVII (1913), 248–54.

[3] W. M. Fontaine, *U.S. Geol. Surv., Mono. 6*, 1883.

[4] J. S. Newberry, *U.S. Geol. Surv., Mono. 14*, 1888.

[5] E. Haycock, "Records of Post-Triassic Changes in Kings County, Nova Scotia," *Trans. N.S. Inst. Sci.*, X (1900), 287–302.

In fragments of shale found in the drift near Kentville, by Haycock, are impressions of *Estheria* of two slightly different types, both of which must be provisionally called *E. ovata*. The shale fragments were evidently derived from the Blomidon shale, and they are very similar to some of the hard barren shale exposed at Rossway.

The paleontological and paleobotanical evidence proves that the Acadian area is a part of the Newark system, and further shows a pronounced similarity between the Newark and the Lettenkohle of Germany.

[*To be continued*]

NOTES ON RIPPLE MARKS

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In a paper on ripple marks, recently published in the *Journal of Geology*, by Dr. E. M. Kindle, the opinion is expressed that the size of ripple marks may bear some relation to the depth of the water in which they were formed. Entertaining the same idea, I have on various occasions taken notes on the size of ripple marks. That most ripple marks vary in size with depth of the water seems to me hardly to admit of a doubt. Ripple marks from 3 to 4 inches in width appear to be most common. They are often to be seen in thoroughly sorted beach sands of all ages, from the Cambrian up to the Pleistocene.

In the Lower Comanchean, in Pecos County, in Texas, I have found some ripple marks of very small size, the smallest I have seen, with one exception. These were noted at several points in some thin-bedded layers of sandstone of fine texture. These sandy layers are interbedded with clays and limestones. A piece of this ripple-bedded rock is shown in Fig. 1, in natural size. Twelve ripples measure together 3 inches across, making an average of one-fourth inch for each ripple, from crest to crest. The depth of the troughs measures about one twenty-fifth of an inch. These ripple marks are symmetrical. A rough mechanical analysis of the sand in this rock is as follows:

Diameter of Grains in Millimeters	Percentages by Weight
1/8 –1/16	80
1/16–1/32	20

Two years ago I found ripple marks of the same size, or possibly slightly smaller, forming in some fine sandy silt in the Rio Grande, in Webb County. The silt had been washed up on some large blocks of sandstone, which were strewn in the channel of the river. It lay in shallow depressions on these rocks, and the water covered

the ripple marks from a half to one inch deep. The wind stirred the surface of the water gently into small waves, and the ripple marks in the sand were seen to be building, under the influence of these waves.

Fig. 2 shows some ripple marks in a fine silty sand of the marine Jurassic, near Minnekahta, South Dakota. They measure $1\frac{1}{3}$ inches from crest to crest and have an average depth of fifteen-hundredths

FIG. 1.—Ripple marks in Comanchean sandstone, from Pecos County, Texas. Natural size.

of an inch. These ripple marks are unsymmetrical, their longer slopes bearing the average ratio of 152 to 100, to the shorter slopes. A mechanical analysis of the sand in this rock was found to be, roughly, as follows:

Diameter of Grains in Millimeters	Percentages by Weight
1/2 –1/4	Trace
1/4 –1/8	55
1/8 –1/16	30
1/16–1/32	15

Some large-sized ripple marks occur in the Ordovician dolomites at Utica, in Illinois. In the old entries where cement rock long ago

was mined for the Utica Cement Works, some ripples have been disclosed that measure from 4 to 5 feet across from crest to crest. This is in a somewhat thin-bedded dolomite, which contains some sand. Evidently this limestone was not a shallow-water deposit. The ripple-bedded layers lie some 100 feet below the base of the St. Peter sandstone.

The widest ripple marks that have come under my observation are in a crinoidal limestone in the lower part of the Burlington,

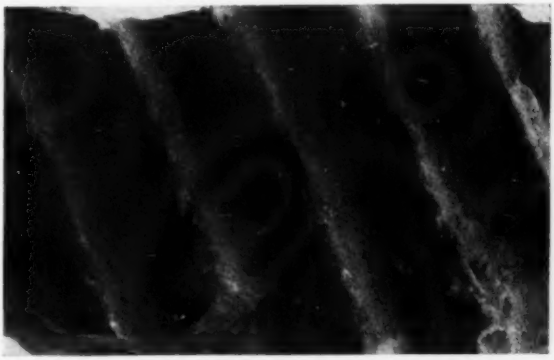

FIG. 2.—Ripple marks on Jurassic sandstone, near Minnekahta, South Dakota. One-half natural size.

in the southeast part of Louisa County, in Iowa. These ripples measure nearly 6 feet from crest to crest, and are at least 6 inches deep. The presence of crinoidal remains in this rock, which contains some shaly material, indicates, if not proves, deep-water conditions. How deep?

Higher up in the geological column I have seen some quite large ripple marks in the Comanchean, in Texas, in a horizon near the Kiamitia clay. About 17 miles west-southwest from Kerrville such ripple marks occur in the bed of Guadalupe River (see Fig. 3). They measure about 14 inches across and are about 1½ inches deep. They are slightly unsymmetrical. The rock in this case is a mixture of calcareous and shaly material, which contains variable quantities of fine sand, so that some layers might more properly be called

sandstone. The same horizon is exposed in the bottom of Bosque River, at Clifton, 155 miles northeast from the locality just mentioned, and again some 6 miles north of Clifton in the same beds in the same stream. Some layers of limestone here show ripple marks that measure 4 feet across, near Clifton (see Fig. 4), and from 2 to 3 feet across at the northernmost locality (Fig. 5). The lime-

FIG. 3.—Ripple marks in thin-bedded sandy limestone in the bottom of Guadalupe River, about 17 miles southwest of Kerrville, Texas.

stone layers here are compact and quite pure in composition, but are interbedded with marly shales.

Perhaps it may be permitted to submit some general remarks anent the phenomena of ripple marks. They shall be brief. Ripple marks must be due to rhythmic variations in currents in the medium of sedimentation. They are in this respect kin to wavelike etchings, known to be caused by rhythmic movements of corrading currents. Perfectly symmetric ripple marks are probably the result of to-and-fro movements of equal extent in both directions, when these movements are such that the velocity of the motion happens to be

sufficiently strong to move material of the coarseness present where the rhythmic motion prevails.

On the bottom of any billowy water, sufficiently shallow for the size of the waves, there must be a to-and-fro motion for each passing wave. For waves of the same size, the deeper the water the more slow and the more limited will this motion be. Hence the less will

FIG. 4.—Ripple marks in Comanchean limestone in the right bank of Bosque River, near Clifton, Texas.

be the diameter of the particles it will be able to stir. There must be a certain depth where the motion will be just speedy enough to stir particles of silt. Where the bottom lies at this depth, and where it is covered with silt, ripple marks will form. Should not their width be determined by the extent of the to-and-fro movement in each direction? This decreases downward according to a known law.

It is evident that the velocity of each to-and-fro movement on the bottom of an agitated body of water begins with zero, rises to a

momentary maximum, and falls again to zero. For particles of different sizes, there must be different times of duration of speeds attaining and exceeding the respective limits effective for their transportation. This time, and hence the latitude of this effective translatory motion, will increase with the fineness of the stirred sediment. With waves of one and the same size, and with the

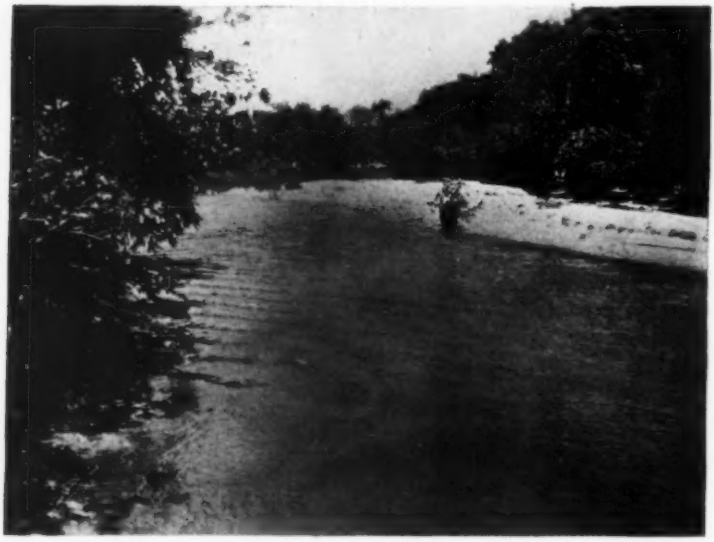

FIG. 5.—Ripple marks in Comanchean limestone in the bed of Bosque River, about 6 miles north of Clifton, Texas.

same depth of water, the width of ripple marks should be greater in fine sediments than in coarse. The currents producing them will carry fine elements farther than coarse. With waves of the same size ripple-mark building in sand should then also take place in somewhat more shallow water than ripple-mark building in silt.

Some ripple marks must be produced by a wavelike or rhythmic motion which results from a reaction by the transported material on translatory bottom currents in water and in the air. No surface

billows in the atmosphere can have anything to do with ripple marks in dune sands. Do dune-sand ripple marks vary in size with wind velocity and coarseness of the sand? They do not vary very much. May ripple marks be formed by a like reaction with bottom currents in deep water? If so, their variation in size may also be small. Such ripple marks, like those in sand dunes, should always be unsymmetrical. Their sizes are probably independent of depth of the water.

PYROPHYLLITIZATION, PINITIZATION, AND SILICIFICATION OF ROCKS AROUND CONCEPTION BAY, NEWFOUNDLAND

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INTRODUCTION

This paper embodies the results of a study of the regional and local alterations which have affected a series of volcanic rocks lying around the borders of the head of Conception Bay, on the Avalon Peninsula of Newfoundland. The field work was accomplished by the writer as a member of the Princeton Geological Expedition in Newfoundland during the summers of 1913 and 1914, in connection with a general study of the pre-Cambrian rocks of this region. The writer is indebted to the Geology Department of Princeton University for facilities for studying these rocks in the field and laboratory; to Dr. C. H. Smyth for supervision in the preparation of the report; to Professor G. Van Ingen for the photographs with which this paper is illustrated and for his interest in the work. The numbers used in this report refer to specimens deposited in the museum of Princeton University.

LOCATION

Regional alterations, such as silicification and chloritization, have affected the volcanic series wherever they outcrop, either in the area here under consideration (Fig. 1), or at other points to the north, such as Clarenville on Trinity Bay or Goose Arm on Bonavista Bay. The local alterations, comprising pyrophyllitization and pinitization, have affected the volcanics only in limited areas; the former being exhibited in a long narrow strip of rocks south of Manuels and the latter in outcrops to the north of the monzonite stock at Woodfords and in minute amounts associated with the pyrophyllite rocks at Manuels.

STRUCTURE

The volcanic series forms the lowest member of the Algonkian rocks in this region and has been mapped as Huronian by Howley (1907). The volcanics at the head of Conception Bay outcrop on

FIG. 1.

the core of a major anticline, the eastern limb of which has been intruded by a huge batholith of granite. On the west side of Holyrood Bay they are also intruded by a stock of monzonite. The rocks are in addition excessively disturbed by profound and intensive faulting, and usually dip steeply.

There is a strong probability that the line of contact between the granite and volcanics at Manuels marks the approximate locus

of a fault zone. For at talc prospects 3, 4, and 5 the volcanics are faulted against either granite or green slate beds, and at prospect 3 the granite adjacent to the fault plane is silicified and pyritized and the rhyolite is silicified and carries traces of pyrophyllite. The pyrophyllite veins in turn are offset by small cross-faults.

CHARACTER OF VOLCANIC SERIES

The volcanics comprise a thick series of rhyolite and basalt flows with corresponding interbedded breccias, crystal tuffs, and tuffs, and a minor amount of waterworn material. The evidence with respect to their origin all points to their having accumulated under subaerial conditions.

TOPOGRAPHY

The topography developed on the volcanics at Manuels is that of a long, narrow, more or less barren plateau about 600 feet above sea-level. The volcanics at the head of Conception Bay are carved into a series of rugged isolated hills or ridges with differential elevations of from 200 to 1,000 feet. Glaciation during the Pleistocene period had a marked effect on the superficial features of the country, and many of the outcrops were scraped and polished by this agency.

WALL ROCKS OF THE PYROPHYLLITE VEINS

The pyrophyllite is confined almost exclusively to the rhyolite flows. Occasionally, however, pockets are found in the rhyolite breccias and conglomerates, but none at all occurs in any of the other rocks. The rhyolite flows exhibit three characteristic structures: flow or banded, spherulitic, and elliptical or lenticular. The spherulites may range in size from micro-spherulites visible only with the high powers of the microscope to huge spheroids as big as a man's head or even larger. They are usually more or less replaced by quartz of chalcedony. The elliptical structure has been called such because of its appearance on the weathered surface of the rock, where it shows as an assemblage of rude ellipses, or as lenses surrounded by a more or less schistose material which may be pyrophyllitized (Fig. 2). The ellipses vary from several inches to a foot in the direction of their longest axis. At talc prospect 5 a rhyolite showing this structure also contains scattered

spherulites. This structure is characteristic of the white rhyolites of this area and may owe its origin to primary flowage phenomena, or to secondary dynamic forces, or probably to the former accentuated by the latter.

DESCRIPTION OF PYROPHYLLITE

The pyrophyllite veins are of such an extent that they attracted attention as a source of talc; many prospects were opened in the

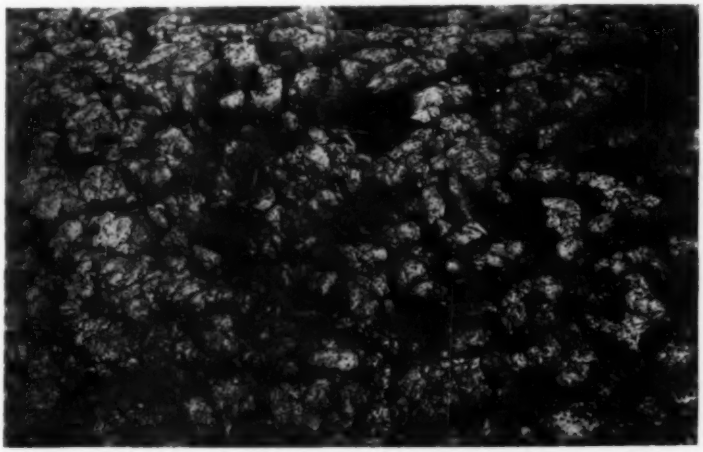


FIG. 2.—Lenticular structure in rhyolite. The material surrounding the more massive portions is partially pyrophyllitized.

deposits and a 2½-mile aerial tram was built to the nearest mine. But, owing probably to the difficulty in separating the pyrophyllite from the admixed quartzose nodules, all work has been abandoned since 1904.

The pyrophyllite where it replaces rhyolite flows, as it does almost exclusively, is a soft cryptocrystalline, light greenish-yellow rock with a waxy luster and a good cleavage parallel to the schistosity. In one case where it replaces the matrix of a volcanic conglomerate it is a light brown and in another where it replaces the matrix of a volcanic breccia it is cream colored.

The pyrophyllite may occur either as single well-defined veins, or as a series of veins, pockets, and lenticels, which together constitute what may be called a pyrophyllitic zone.

The former character is illustrated at talc prospect 5, where the pyrophyllite forms a vein about 500 feet long and varying from 6 to 15 feet in width in a white, densely spherulitic rhyolite. Near the one end of the vein which is exposed the pyrophyllite is full of nodules and stringers of the rhyolite, but becomes almost clear pyrophyllite in its central portion.

The latter character (pyrophyllitic zone) may be illustrated by the character of the pyrophyllite deposits at talc prospects 1 and 2. The country rock of the pyrophyllitic zones may be so altered as to constitute a quartz-pyrophyllite schist consisting of microcrystalline quartz and pyrophyllite, as at talc prospect 1, or it may be partially pyrophyllitized, as at talc prospect 2, or relatively unaltered as at talc prospect 4.

At talc prospect 1 large masses of pyrophyllite occur in pockets from 1 to 15 feet in diameter containing more or less country rock, or as thin sheets incasing lenses of quartz-pyrophyllite rock oriented parallel to the cleavage. It frequently occurs as an interlacing network of films veneering lenses of the quartz-pyrophyllite rock or as lenticels replacing the matrix between adjacent quartzose nodules. The pyrophyllite (*222 E 2 f x*) usually serves simply as a matrix for these nodules varying from a fraction of an inch to several feet in diameter, and even hand specimens are infrequent which do not contain one or more of them. Ramifying stringers of country rock may wander aimlessly through the pyrophyllite (Fig. 3) and veins of pyrophyllite reticulate in the country rock. It is quite possible that the nodular structure originated through the total replacement by pyrophyllite of the sheared zones between lenses of a rhyolite like that shown in Fig. 2, and the alteration of the lenses themselves to a quartz-pyrophyllite rock. On the west side of the talc mine here small pockets of cream-colored pyrophyllite, weathering green or yellow, are found replacing the matrix of a very coarse rhyolite breccia.

At talc prospect 2 there is a pyrophyllite zone about 30 feet wide in which pyrophyllite constitutes from a small percentage

to one-half of the rock. Paralleling this zone at a distance of about 20 feet is a pyrophyllite vein 18 feet in width with a 3-foot stringer and many spheroids of altered rhyolite wall rock in its central portion. The spheroids vary from an inch to a foot in diameter, but average about 4 inches. A lenticular or elliptical structure (Fig. 2) characterizes the rhyolite adjacent to the vein and the schistose matrix of the lenses or ellipsoids is partially pyrophyllitized. A few hundred feet north of here a light brownish pyrophyllite is

FIG. 3.—View of portion of pyrophyllite vein, showing intermingling of pyrophyllite and country rock. p = pyrophyllite; q = quartz-pyrophyllite.

found replacing portions of the matrix of a white rhyolite conglomerate.

DESCRIPTION OF PINITE

At Manuels the pinite is a relatively rare constituent and is interesting only from the viewpoint of its origin. It is best exhibited at talc prospect 5 and at a point marked 228 *D* 1 on the map. Fig. 4 is a photograph taken at this latter locality and represents the matrix of a spherulitic rhyolite (228 *D* 1 *h*) replaced by dark-colored pinite. The spherulites here average about 1 inch

in diameter and are so intermingled with smaller ones as to make up almost the entire bulk of the rock. The pinitized groundmass possesses a waxy luster, dark dirty-green in color, and is quite soft. The parting of the pinite is in general parallel to the cleavage of the rhyolite, although in detail it is a series of curving shell-like scales, owing to its parting following the circumference of the more resistant spherulites. An analysis of this matrix is given in this

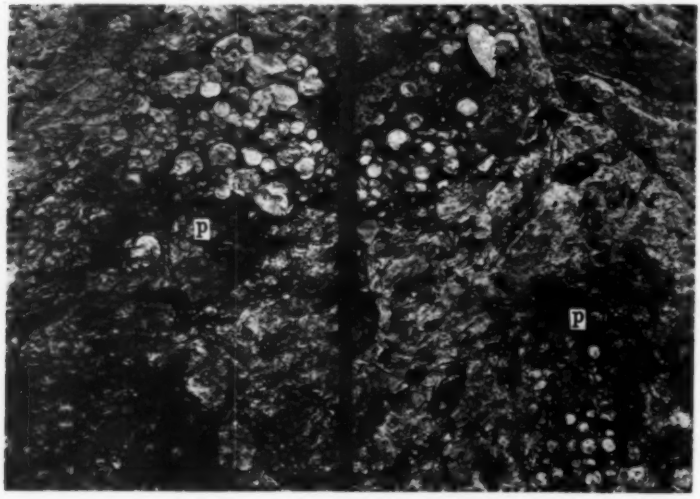


FIG. 4.—Pinite (*p*), replacing portions of the matrix of a spherulitic rhyolite

Journal on p. 137 (No. 7). Very rarely pinite is found along the original contraction cracks of the spherulites or at the heart of a spherulite. Patches, lenticels, and minute veins of pinite are found throughout the rhyolite flows and agglomerates, often replacing the matrix of spherulitic zones (228 *G* 1 *k*) or certain flow lines (222 *E* 2 *x*).

In the valley of Harbour Main Brook, among a series of rhyolite flows and tuffs, tuff beds up to 75 feet thick have been partially altered to pinite, and spherulitic rhyolite flows up to 30 feet thick are streaked and banded with pinite. These rocks have been

prospected to depths of 15 and 20 feet, probably under the misapprehension that they carried pyrophyllite. A chemical analysis of the pinitized groundmass of the spherulitic rhyolite is given on p. 137 (No. 8).

CHEMICAL ANALYSES

Chemical analyses of eight typical rocks showing the various types and some of the stages of the alterations and replacements which have affected the volcanics are here given. These analyses were recalculated to correspond approximately to the mineral composition of the rocks. Small amounts of water of absorption, iron oxides except in No. 7, excess alumina, etc., have been lumped together as such under "other constituents." This involves of course a slight but inappreciable error in the proportions of the other minerals. It is probable that some sericite is also present in rocks Nos. 3 and 4, but owing to the difficulty of distinguishing sericite from pyrophyllite under the microscope, and because of the fact that the decrease in potash with an increase in water

TABLE I

SHOWING CHARACTER OF ALTERATIONS OF RHYOLITE

	1	2	3	4	5	6	7	8
SiO_2	76.24	80.60	74.51	72.10	65.04	88.09	54.47	61.07
Al_2O_3	13.94	11.27	17.12	21.51	29.49	9.53	27.14	22.99
Fe_2O_3	0.89	0.89	1.28	0.54	0.28	0.20	2.58	1.56
FeO	0.13	0.08	0.08	n.d.	n.d.	n.d.	0.47	0.42
MgO	0.27	0.31	0.04	0.04	0.04	0.05	2.44	0.65
CaO	1.07	0.66	0.13	0.49	0.10	0.58	0.81	1.63
Na_2O	2.55	1.16	0.48	0.42	0.33	n.d.	0.68	0.63
K_2O	4.95	4.68	3.68	2.21	0.33	n.d.	8.01	7.58
H_2O+	0.15	0.49	2.44	3.09	4.84	1.68	3.44	2.93
H_2O-	0.03	0.11	0.09	0.23	0.03	0.04	0.27	0.23
MnO	Trace	Trace	Trace	n.d.	n.d.	n.d.	0.09	0.17
	100.22	100.25	99.85	100.63	100.48	100.17	100.40	99.86
Sp. G.	2.64	2.64	2.73	2.77	2.83	2.70	2.82	2.70

1. Unaltered dark-gray flow rhyolite (228 *D* 1 *c*).
2. Silicified drab spherulitic rhyolite (228 *F* 2).
3. Pyrophyllitized rhyolite Matrix of lenses; talc prospect 2 (228 *F* 3).
4. Pyrophyllitized rhyolite. Matrix of spherulite; talc prospect 1 (222 *E* 2 *j*).
5. Pyrophyllite. Light greenish-yellow waxy pyrophyllite from talc prospect 1 (222 *E* 2 *f*).
6. Quartz-pyrophyllite schist. Nodule in pyrophyllite (5); talc prospect 1 (222 *E* 2 *g*).
7. Pinite; dark dirty-green waxy, matrix of spherulitic rhyolite. Manuels (228 *D* 1 *k*).
8. Pinite schist. Matrix of spherulitic rhyolite, Harbour Main (232 *E* 2 *b*).

indicates that the replacing mineral is pyrophyllite, it has been calculated as that alone. Owing also to the liability of error involved in assigning the elements of the pinitic rocks to the correct minerals in the right proportions, only a rough estimate of their mineral composition is given. The iron oxides and magnesia of No. 7 must be present as an integral part of the white mica molecule, as no other mineral except quartz can be distinguished in thin section.

TABLE II

RECALCULATED ANALYSES

	Quartz	Orthoclase	Albite	Anorthite	Pyrophyllite	Other Constituents	White Mica
1.....	39.6	29.4	21.4	5.2	.....	4.4	.....
2.....	53.7	27.8	9.9	3.1	.....	5.5	.....
3.....	28.3	21.7	4.2	0.6	43.6	1.6	.....
4.....	18.7	13.3	3.6	2.5	61.1	0.8	.....
5.....	1.3	.....	.....	.....	91.8	1.6	5.3
6.....	65.7	.....	.....	.....	33.4	0.9	.....
7.....	Present	.....	.....	.....	.....	.....	About 75 per cent
8.....	Present	Present	Present	.....	.....	.....	About 60 per cent

PETROGRAPHY

No. 1 (228 *D* 1 *c*). This specimen was taken near the top of a 50-foot banded reddish-gray felsite flow. In thin section the texture varies from microfelsitic to very minutely microcrystalline and the flow lines are marked by hematite dust. The flow lines are sharply curved and crenulated and several are replaced by quartz, especially in the loops of the curves, so that the rock analyzed as representing the composition of the original rhyolite only approximates such an unaltered condition.

No. 2 (228 *F* 2). This rock is a drab to fawn-colored microspherulitic rhyolite with secondary iron oxide in veinlets and specks. In thin section the groundmass is a finely microcrystalline aggregate of quartz and orthoclase, with fan-shaped microspherulitic areas. Secondary quartz is present as grains and lenses, as well as replacing portions of the spherulitic aggregates. A minute amount of sericite and quartz occurs along fractures.

No. 3 (*228 F 3*). This specimen, a pyrophyllitized rhyolite, was taken from the slightly sheared matrix surrounding lenses of white rhyolite (Fig. 2) adjacent to pyrophyllite veins at talc prospect 2. In thin section the rock shows as a microcrystalline aggregate of granular quartz and feldspar and of scales and fibers of pyrophyllite in about equal amounts.

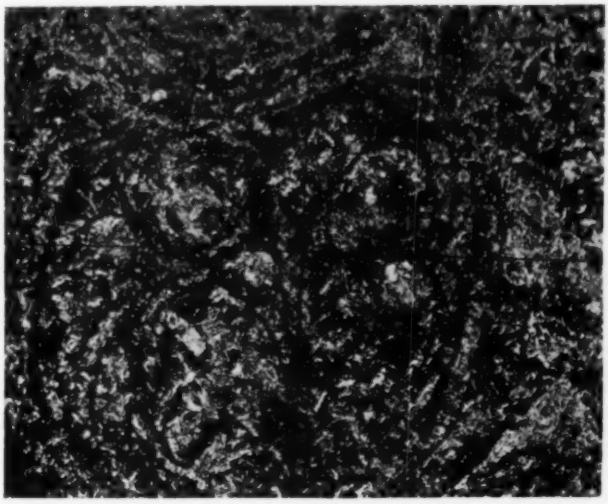


FIG. 5.—Perlitic structure preserved in pyrophyllitized rhyolite. Ordinary light, ×60.

No. 4 (*222 E 2 j*). This is the matrix, a pyrophyllitized rhyolite, in which a 6-inch spherulite was found essentially unaltered. In thin section the rock consists of an aggregate of very minute microscopic scales of pyrophyllite, complete except for a remarkably well-preserved perlitic structure, outlined by microcrystalline quartz with probably some orthoclase (Fig. 5). The rock is in a much more advanced stage of alteration to pyrophyllite than the preceding specimen.

No. 5 (*222 E 2 f*). Light greenish-yellow pyrophyllite with a fair cleavage. In thin section the rock is seen to be composed of

a homogeneous felt of exceeding minute microscopic scales and fibers of pyrophyllite, with a strong tendency toward a very good alignment parallel to the cleavage in a section at right angles to it and with long, parallel, fibrous shreds in a section approximately parallel to the cleavage.

No. 6 (*222 E 2 g*). A white quartzose nodule or lense of quartz-pyrophyllite schist about 1 foot in diameter taken from a pyrophyllite vein at talc prospect 1. In thin section the rock presents what might be called a micro-blotchy groundmass composed of aggregates of either microcrystalline quartz or of scales of pyrophyllite. Some fibers of pyrophyllite also occur interstitially in the quartz areas.

No. 7 (*228 D 1 h*). This is the dark grayish-olive, waxy-lustered matrix of the spherulitic rhyolite illustrated in Fig. 4. In thin section the rock is seen to consist of an aggregate of extremely fine shreds and scales of white mica with lines of partially replaced microcrystalline quartz which are probably replacements of certain of the original rhyolite flow lines not yet entirely replaced by the white mica.

No. 8 (*232 E 2 b*). Grayish-olive pinite schist with small unaltered spherulites or spherulites partially replaced by quartz. In thin section the material appears as a perlitic microcrystalline groundmass of quartz and orthoclase partially replaced by sericite. The perlitic cracks are outlined by threads of sericite fibers, as illustrated in Fig. 6, and they often form the boundaries of serocitized areas which present the appearance of eyes, sometimes with a reticulating network of sericite veins connecting two adjacent eyes. Within the sericitic material, isolated microspherulites, clusters of microspherulites, and long axiolites are often preserved intact. A few phenocrysts of orthoclase are present and are remarkably fresh, although occasionally flecked with sericite. The secondary material consists of an aggregate of microscopic sericite scales and fibers associated with grains and areas of secondary quartz. Minerals originating through decomposition at the surface are completely absent except for a trace of iron oxides in the groundmass and a slight cloudiness in the feldspars, probably due to kaolin.

PRELIMINARY SILICIFICATION

The first process of alteration which operated on the Avondale Volcanics was that of regional silicification. On these preliminary silicified rocks a local series of alterations, those of pyrophyllitization, pinitization, and further local silicification, were superimposed.

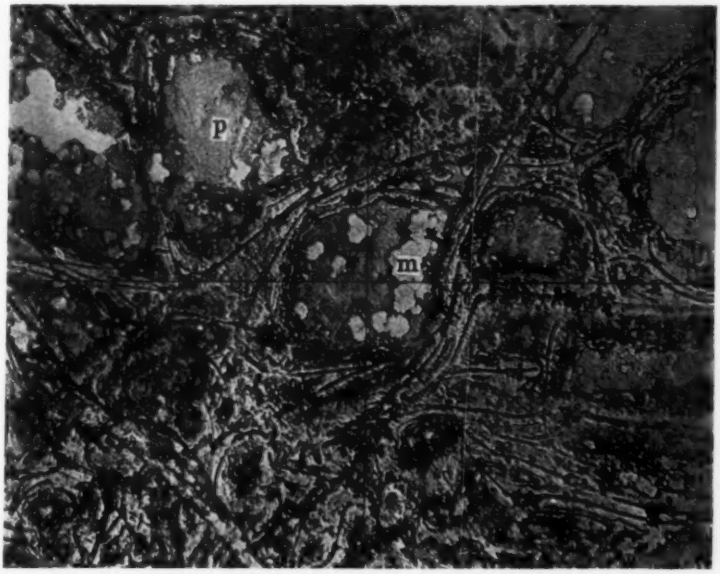


FIG. 6.—Perlitic structure preserved in pinitized rhyolite. Ordinary light, ×35. *m* = microspherulites; *p* = pinite.

This is evidenced by the following data: (1) few later quartz veins are found traversing the pinite, quartz schists, or pyrophyllite; (2) fragments of breccia in the volcanic breccias often exhibit quartz veins which stop abruptly at the contact with the matrix, and (3) under the microscope aggregates of sericite scales are found replacing granular quartz which had previously replaced the heart of a spherulite, and these sericite scales finger into and inclose unreplaced fragments of quartz, proving definitely their later origin.

The spherulitic rhyolites are the rocks which exhibit most clearly the manner in which the silicification has taken place. The silica here is present for the greater part as a milky-white chalcedonic quartz, but vitreous, granular, and white vein-like quartz as well as quartz crystals are common. The chalcedony usually forms the outer borders of the concentric crescent-shaped areas, and of the hearts of the replaced zones of the spherulites, while the inner portion may be recrystallized to form comb structure through the interlocking of quartz crystals, or little geodes with terminated crystals projecting into a small cavity, or granular vitreous quartz. When only one form of the silica is present it is very generally of a chalcedonic nature. It is interesting to note that while at Manuels it is the spherulites of the rhyolites which are most generally replaced, at Clarenville it was the groundmass which was replaced instead of the spherulites, because of the perlitic structure of the former offering the most favorable surfaces for attack. The banded rhyolites are often lined or streaked with quartz veins parallel to the planes of flow, which in some cases are a result of replacement and in others of vein filling. In thin section, lenses, lines, and granules of secondary quartz are found to be a common characteristic of the slightly silicified banded flows. Fig. 7 illustrates the preservation of the perlitic structure in the quartz which is replacing the groundmass of a spherulitic rhyolite from Clarenville. The perlitic cracks are outlined by sericite.

That the silicification of the rhyolites has been due to secondary metasomatic processes and is not a primary phenomenon is indicated by (1) the interruption of fluxion lines by the replacing quartz, (2) by the presence of unsupported fragments of unreplaced rhyolite in the quartz areas, and (3) by the preservation in the quartz of original structures of the rhyolite, such as the perlitic structure.

The first stage then in the alteration of these volcanics has consisted in the silicification under relatively static conditions by hot siliceous waters of rhyolite flows which may be represented as having had a similar chemical composition to the present relatively unaltered gray felsites. Analysis No. 1 may be taken as the com-

position of the original rock from which the silicified rhyolite represented by analysis No. 2 has been derived.

Chemically this has resulted in a decrease in the percentages of potash, soda, lime, and alumina, an increase in the percentage of silica, and a relative decrease in the percentage of sodium with respect to potassium. The process operated through the replacement of the feldspars by quartz and a relatively more rapid

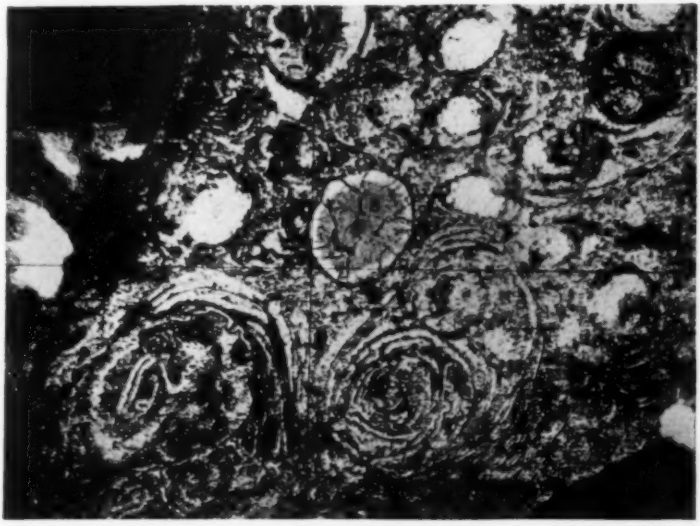


FIG. 7.—Perlitic structure preserved in quartz replacing the groundmass of a spherulitic rhyolite. Ordinary light, ×35. *s* = spherulite.

replacement of the soda feldspars than of the potash feldspars. The solutions which effected this alteration doubtless belonged to the same general period of volcanic activity as the extrusion of the lavas themselves.

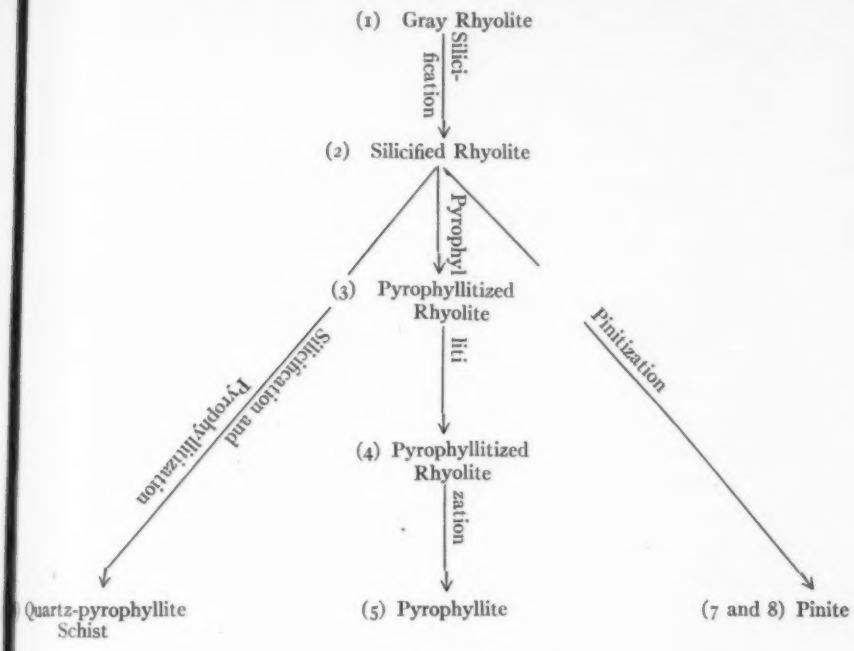
PYROPHYLLITIZATION, PINITIZATION, AND SILICIFICATION

As has been remarked before, a later series of local alteration processes has been superimposed on the already widespread slightly silicified volcanics. The origins of the pyrophyllite, the pinite, and

the quartz-pyrophyllite or quartz schists are so intimately interwoven that all may be treated together.

The proof that these rocks have originated through replacement is based on the following data: (1) the preservation of the structures of the primary rock in the secondary rock, (2) the presence of unattached and unsupported portions of the country rock within the replacement products, (3) the introduction of large quantities of some elements and the solution of others without any notable change in volume or porosity, (4) gradational contacts, and (5) the massive homogeneity of all the rocks, and especially of the pyrophyllite, which does not show the foliated crystalline structure so characteristic of pyrophyllite veins which fill preexisting fractures. To quote examples which belong to the first category, we find the following structures preserved in pyrophyllitized rhyolite: (1) flow structure, (2) spherulites, (3) pebbles of a partially replaced conglomerate, and (4) perlitic structure (Fig. 5); these in the quartz-pyrophyllite rocks: (1) spherulites and (2) breccia structure; the following in pyrophyllite: (1) fragments of volcanic breccias, (2) pebbles of conglomerates, and (3) spherulites; while in the pinite and pinite schists we have preserved spherulites, traces of flow structure, axiolites, microspherulites, and perlitic structure (Fig. 6). Additional evidence of replacement is found in the inclusions and stringers of country rock within the pyrophyllite veins and the intimate manner in which the two are often intermixed. Not only this most convincing field evidence but chemical considerations prove almost conclusively that these rocks must have originated through replacement of rhyolite or rhyolitic volcanics.

From a study of the foregoing field and chemical evidence, conclusions have been drawn as to the genetic relationships of the eight different rocks described under "Chemical Analyses" and "Petrography," and as to the succession of processes which produced them. This relationship is graphically represented by the following diagram, in which the numbers refer to the chemical analyses given under "Chemical Analyses," which may be taken as typifying the composition of the respective rocks:



TEMPERATURE OF FORMATION OF PYROPHYLLITE

Clapp (1914, 120) assumes that alunite and pyrophyllite are probably developed only under moderate conditions of pressure and temperature such as exist near the surface. Although this is a common mode of origin for both alunite and pyrophyllite, it is certainly not the only set of conditions under which the latter forms.

For instance, pyrophyllite is noted by Dana (1909) as a mineral often forming the base of schists and gneisses, and by Lacroix (1895) as a mineral of the crystalline schists and Paleozoic metamorphics.

Artificially, K. von Chrustchoff (1894) obtained what he believed to be pyrophyllite by heating gelatinous silica, gelatinous alumina, and gelatinous zirconium hydrate in a platinum tube at

increasing temperature for six days. The product obtained was a zirconia-bearing pyrophyllite. Its specific gravity was 2.87 and it appeared in thin hexagonal plates, which he states are not true hexagonal plates unless there be optical anomalies.

SiO_2	Al_2O_3	ZrO_2	H_2O	Total
53.65	23.76	14.54	7.86	99.81

Le Chatelier (1887) determined the points at which pyrophyllite loses its water by noting the points at which the temperature remained constant with absorption of heat and found two such points, the first at 700° and the second at 850°.

From the foregoing data it is evident that pyrophyllite is a mineral which may form under conditions varying from the high temperatures of dynamic metamorphism to the near-surface temperatures and pressures of solfataric agencies.

TEMPERATURE OF FORMATION OF PINITE

Pinite, if considered as an impure sericite, as suggested by Clarke (1911), has a varied range of conditions under which it may form. Clarke states, however, that "the alteration [to sericite] is most conspicuous in regions where dynamic metamorphism has been most intense, high temperature, the chemical activity of water and mechanical stress all working together to bring it about."

A green micaceous mineral described as mariposite by Silliman, and whose composition, shown by two analyses, as suggested by Hillebrand (1895), resembles pinite, is characteristic of the mother lode in Tuolumne and Mariposa counties, California.

Crosby (1880) describes pinite as a product of surface decomposition of petrosilex and felsites in the vicinity of Boston, Massachusetts. Bell (1887) found it at Ballater Pass interspersed through granitic rocks and along their joint planes, and ascribes its origin to the decomposition or alteration of orthoclase feldspar, an intermediate stage in its conversion into kaolin.

Cole (1886) describes pinite occurring as an alteration product of spherulitic rhyolites, in conjunction with silicified spherulites. He suggests that thermal waters are responsible for the origin of both the pinite and quartz.

From these references it is evident either that there is a difference of opinion as to the conditions under which pinite forms, or that it is stable under widely variant temperatures and pressures. It is probable, however, that it demands higher temperatures and pressures than exist at the surface as conditions for its most favorable development, and such is doubtless the case with respect to the pinite of Conception Bay.

ALTERNATIVE DEVELOPMENT OF PINITE OR PYROPHYLLITE

Since sericite or pinite is the usual product of hydrothermal alteration it is pertinent to inquire if any reason can be found why in certain cases pyrophyllite should be the product formed. A possible equation (1) representing the formation of sericite from orthoclase is quoted from Clarke (1911), and a similar possible equation (2) representing the formation of pyrophyllite from orthoclase is given below:

(1) $6\ KAlSi_3O_8 + 2H_2O \rightleftharpoons \underset{\text{(sericite)}}{2KH_2Al_3Si_3O_{12}} + 2K_2SiO_3 + 10SiO_2$

(2) $6\ KAlSi_3O_8 + 3H_2O \rightleftharpoons \underset{\text{(pyrophyllite)}}{6HAlSi_2O_6} + 3K_2SiO + 3SiO_2$

From these equations three factors are suggested as the possible elements influencing the alternative development of sericite and pyrophyllite: (1) the effectiveness of hydrolysis, (2) the mass action of the excess silica in solution, and (3) the mass-action effect of excess potash in solution. The dominance of the first two factors would be conducive to the formation of pyrophyllite, and the dominance of the third factor would be favorable to the production of sericite. This may be illustrated more graphically, without however implying anything as to the actual mode of operation, by writing the equation for the formation of pyrophyllite from sericite as a balanced reaction:

$$\underset{\text{(sericite)}}{2KH_2Al_3Si_3O_{12}} + 6SiO_2 + H_2O \rightleftharpoons \underset{\text{(pyrophyllite)}}{6HAlSi_2O_6} + K_2SiO_3$$

If now the quantity of silica is present in the solution in large enough excess and the effectiveness of hydrolysis is relatively stronger, the reaction will produce pyrophyllite:

$$\underset{\text{(sericite)}}{2KH_2Al_3Si_3O_{12}} + 6SiO_2 + H_2O \rightarrow \underset{\text{(pyrophyllite)}}{6HAlSi_2O_6} + K_2SiO_3$$

while if K_2SiO_3 or potash in some other form is present in large enough quantity the alternative reaction will take place and sericite will be produced:

$$\underset{\text{(pyrophyllite)}}{6HAlSi_2O_6} + K_2SiO_3 \rightarrow \underset{\text{(sericite)}}{2KH_2Al_3Si_3O_{12}} + 6SiO_2 + H_2O$$

CHEMICAL PHENOMENA CONNECTED WITH ORIGIN OF PYROPHYLLITE

From a comparison of the analysis (No. 2) of the silicified rhyolite with that of the pyrophyllite (No. 5), it will be seen that the change in composition has been such as might have been brought about essentially through three processes: (1) the introduction of alumina, (2) the replacement of the alkalies by hydroxyl, and (3) the solution of silica. The analyses 2, 3, 4, and 5, recalculated into their mineral composition, show a direct transition from the country rock (the silicified rhyolite) into pyrophyllite through a decrease in the quantity of quartz, feldspars, and impurities and a simultaneous increase in the content of pyrophyllite. In order that the original rock may be so altered as to give the mineral analyses shown by the transitional rocks, it is necessary that metasomatic replacement of both the quartz and the feldspars should have proceeded synchronously and at a much faster rate with respect to the quartz than with respect to the feldspars. This process would involve the introduction of large amounts of alumina, the gradual replacement of the alkalies by hydroxyl at a more rapid rate in the case of the soda than of the potash, and the solution of portions of both the silica existing in combination with other elements in the rock and that present as free quartz.

The heterogeneous, blotchy character of the quartz-pyrophyllite rock when seen in thin section suggests that the rock may have been in the condition of a more or less homogeneous glass when acted upon by the silicifying and pyrophyllitizing solutions, in view of the fact that its chemical composition and mineral arrangement

would involve the simultaneous replacement of the feldspars of a crystalline rock by silica and pyrophyllite, and of its quartz by pyrophyllite.

ORIGIN OF PINITE AT MANUELS

From a further study of the chemical analyses, it becomes evident that while the silicified rhyolites, pyrophyllitized rhyolites, and pyrophyllite have all decreased in their content of iron, magnesium, potassium, and soda, the pinite analysis (7) shows a decided increase in the first three of these elements. During the formation of pyrophyllite vast quantities of potash must have been liberated and carried in solution in the circulating waters. Is it not possible that away from the main channels these waters deposited their load as pinitic replacements of the rhyolite under the control of lower temperatures and pressures and the mass-action effects of the excess potash in solution? It seems reasonable to suppose that the pinite here was an essentially contemporaneous formation with the quartz-pyrophyllite schists and pyrophyllite, receiving some of the magnesia, potash, and iron released by the formation of the pyrophyllite, as the quartz schists have received some of the silica originating at the same time.

ORIGIN OF THE PINITE SCHIST AT HARBOUR MAIN

The chemical analysis of the pinite schist recalculated for sericite gives the rock a mineral composition of about $\frac{3}{5}$ sericite and $\frac{2}{5}$ quartz, feldspar, and other constituents. Examination of thin sections shows that this result has been brought about through the replacement of both the feldspar and quartz by sericite. If we consider the rock previous to pinitization to have had the composition of the silicified rhyolite (No. 2), then the process cited involves the substitution of potash for soda in the feldspars, the addition of potash and alumina, and the subtraction of silica and soda.

There is considerable evidence that this rock was not formed at the surface. There are no secondary products of decomposition, such as kaolin or limonite, associated with the pinite. It occurs in quantity only in certain zones and extends to a considerable depth, as exposed in a prospect pit for talc, south of Harbour Main.

Where it occurs as patches or specks it bears no apparent relation to the surface. Furthermore, the pinitic material has undergone dynamic metamorphism and is sheared and cleaved, while chemically its origin involves the introduction of alumina and potash through replacement.

GENERAL OBSERVATIONS

It is quite probable that the foregoing processes operated under conditions of some dynamic movement, as their characteristic development is along shear zones, and their products assume a lenticular structure which is as characteristic of the minuter structures of the rocks as of the veins themselves. Moreover, the rocks themselves have been sheared and possess a more or less prevalent cleavage, conditioned by the growth of their constituent minerals in more or less parallel arrangement.

It is probable that the factors determining which of these three rocks—pyrophyllite, pinite, and quartz-pyrophyllite—shall form are to be found in the temperature, pressure, and chemical content of the solutions themselves, in the relative effectiveness of hydrolysis, and in the mass-action effects of the compounds in solution.

The concentration of the potash in the pinite, and the silica in the quartz-pyrophyllite rocks may be accounted for on the theory of a redistribution of the elements, but the tremendous contributions of alumina represented by the pyrophyllite, and to a minor extent by the pinite, must be accounted for otherwise. The close connection between the pyrophyllite deposits and the granite-Avondale Volcanics contact south of Manuels and between the pinitized rhyolites and the Woodfords monzonite stock is hence of significance. It may be that there is no genetic connection between these minerals and the intrusives, and that their formation was entirely dependent on the locus of fault zones in those localities. But there are numerous other profound fault and shear zones in this area with no exceptional alterations.

Hence it seems probable that when faulting took place between the volcanics and the intrusive granite or monzonite, the still hot magmatic waters were released and found their way upward along these fault zones, either contributing the alumina directly, or per-

haps indirectly through mingling with already aluminous solutions. Such solutions, if present, would be those which had accomplished the silicification of the volcanics with their attendant solution of alumina under static conditions and hence before the period of orogenic movement, which folded the volcanics and witnessed the intrusion of the plutonic batholith and stock. The evidence bearing on the exact dates of the periods of faulting and folding, however, is not conclusive.

COMPARISONS WITH OTHER DEPOSITS

Comparisons with other deposits show that, according to the descriptions, the pyrophyllite deposits of the Pambula goldfield, New South Wales, and of Chatham and Moore counties, North Carolina, are essentially similar to those in Newfoundland, and it is here suggested that possibly they have had a similar origin.

Clapp (1914) described quartz-pyrophyllite rocks from Kyuquot Sound, Vancouver Island, which may be taken as a type of pyrophyllite deposits developed by solfataric agencies under conditions of temperature and pressure existing near the surface; while those of Newfoundland are a type originating under intermediate conditions of temperature and pressure.

In the first case the pyrophyllite rocks are associated with alunite and in the latter case with pinite. In the Kyuquot deposits the original quartz of the replaced dacite has not suffered any loss except in one doubtful case, while the distinctive feature of the Newfoundland rocks has been the replacement of quartz in rhyolites by pyrophyllite.

The two deposits are similar in that, in both cases, the rocks are associated with intrusive batholiths, the one with a feldspathic quartz diorite and the other with granite. Both are metasomatic replacements of acid volcanics, while in the zone of alteration there seems to have been some transfer of material, and soda, lime, magnesia, and iron oxides have been lost in each case.

CONCLUSION

From the foregoing evidence the conclusion may be drawn that: the pyrophyllite, pinite, and quartz-pyrophyllite schists of the

Avondale Volcanics owe their formation to metasomatic replacement and alteration of previously silicified rhyolites or rhyolitic volcanics by thermal waters, under conditions of dynamic stress and intermediate temperatures and pressures, operating along channels primarily determined by fault or shear zones. Chemically, the salient features of these alterations have been the introduction of alumina, the more or less complete substitution of the hydroxyl element in place of the alkalies, and the solution of soda. The solutions instrumental in causing these alterations may have been to a greater or less extent juvenile waters emanating from the intrusive granite batholith and monzonite stock at some period subsequent to the time of their injection.

May, 1915

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THE ORIGIN OF RED BEDS

A STUDY OF THE CONDITIONS OF ORIGIN OF THE PERMO-CARBONIFEROUS AND TRIASSIC RED BEDS OF THE WESTERN UNITED STATES[1]

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WAS THE COLORING MATTER A CHEMICAL OR A MECHANICAL SEDIMENT?

[1] A revision of a thesis presented in partial fulfilment of the requirements for the degree of Master of Arts at the University of Wisconsin, in June, 1914.

PART II

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SUMMARY

PART I

PREFACE

Certain general facts concerning the origin of the western Red Beds have been known for some time and have become incorporated into current textbooks. There has been, however, much difference of opinion in the interpretation of some features of this remarkable group of sediments, especially as to the significance of the color itself. This paper is devoted to an investigation of the causes and history of the coloring matter, which, more than all other features put together, distinguishes the Red Beds from other sedimentary series.

The investigation on which this paper is based has been chiefly a study of the literature to which reference is made in the footnotes, together with all available published descriptions of the western Red Beds, and much miscellaneous literature dealing with related subjects. The writer's first-hand acquaintance with the Red Beds has been gained from two summers of field work in Wyoming and Idaho (under the direction of Eliot Blackwelder, of the United

States Geological Survey), and from laboratory study of thin sections. For valuable suggestions and criticism and for review of the manuscript the writer is indebted to Messrs. Eliot Blackwelder, A. N. Winchell, W. J. Mead, and F. T. Thwaites, of the University of Wisconsin; to Messrs. W. H. Emmons, C. R. Stauffer, F. F. Grout, C. J. Posey, and A. W. Johnston, of the University of Minnesota; and to Dr. R. D. Salisbury, of the University of Chicago.

SUMMARY DESCRIPTION OF THE FEATURES CONSIDERED

List of formations.—Clastic sedimentary strata of reddish color outcrop or constitute the uppermost part of the bedrock over about 4 per cent of the area of the United States and exist beneath a cover of younger sediments under an additional area probably twice as large. Much the greatest volume of such strata is included in the single group of rocks which forms the subject of the present study: namely, those Red Beds which outcrop in many areas from Kansas and Texas to Arizona and Montana (Fig. 1). The formations included in this group are related closely in age, ranging from Pennsylvanian to Triassic, or possibly Jurassic, and are probably for the most part physically continuous. They comprise the Cimarron series of Kansas and Oklahoma, and the Wichita, Clear Fork, Double Mountain, Greer, Quartermaster, and Dockum beds of Texas; the Wyoming, Fountain, and Maroon formations of central Colorado; the Cutler and Dolores of southwestern Colorado; the Aubrey, Shinarump, Vermilion Cliff, and Moencopie of the Colorado Plateau, with the Saliferous and Zuni of the Zuni Plateau in New Mexico; the Spearfish, Opeche, and Chugwater of Wyoming and southern Montana; and the Ankareh and Nugget of southeastern Idaho and northeastern Utah.

Similar Red Beds constitute a large part of the sediments of the Newark series (Triassic) of the Appalachian Piedmont region—typified by the Stockton and Brunswick formations of New Jersey.[1] They make up also, among others, much of the Catskill formation (Devonian) of eastern New York and Pennsylvania; the Medina and Clinton (Silurian) of New York; the Vernon shale (Silurian, a

[1] See H. B. Kümmel, *Ann. Rept. of the State Geologist of New Jersey*, 1896.

FIG. 1.—Map showing distribution of the western Red Beds

member of the Salina beds), the Bedford shale (Mississippian), and the Dunkard series (Permian) of Ohio; the Lake Superior sandstone (probably Algonkian) of northern Michigan, Wisconsin, and eastern Minnesota; and the Belt series (Algonkian) of Montana and British Columbia. Occasional reference will be made in this paper to these formations, and to Red Beds in other countries as well; but this discussion applies especially to the western group, with which the writer is most familiar.

Colors of gypsum and limestone.—The beds of gypsum occurring in many areas of Red Beds strata are described everywhere as remarkably pure and white, except where stained by a red coating washed down from overlying clastic sediments. The same is true of a majority of the limestones and dolomites occurring in the red series, most of which are gray or drab or bluish on fresh fracture. Even the famous Redwall limestone (2,500 feet thick) underlying the Aubrey Red Beds of the Grand Canyon section is gray on fresh fracture;[1] its surface color is due to concentration of iron oxide by weathering, to wash from above, or to both of these causes. Exceptions are found in limestone or dolomite bands in the lower Chugwater in Wyoming,[2] and in the Minnekahta limestone[3] of the Hartville quadrangle, which are characteristically purplish or rosy gray; and locally in certain limestones in northern Oklahoma, where they are in transition to sandstone.[4] These limestones are in most localities nearly or quite barren of fossils.

Color in clastic strata.—Another and perhaps a yet more significant fact is the variation of hue among the clastic strata. Interbedding of greenish, gray, or buff beds with red sediments is found in a majority of Red Beds sections; notably in the Saliferous of the Colorado and Zuni plateaus, in the Dockum group of Texas, and in

FIG. 1.—Map showing distribution of the western Red Beds

[1] G. K. Gilbert, "Geology of Portions of Nevada, Utah, California, and Arizona," *U.S. Geog. Surveys W. of the 100th Meridian*, III (1875), 177–78.

[2] Eliot Blackwelder and C. W. Tomlinson, "Field Notes on Work in Western Wyoming, 1910 and 1911," unpublished; property of the United States Geological Survey.

[3] W. S. T. Smith, Hartville Folio (No. 91), *Geol. Atlas of the U.S., U.S. Geol. Survey*, 1903.

[4] J. W. Beede, "The Neva Limestone in Northern Oklahoma," *Okla. Geol. Survey Bull. No. 21*, 1914, p. 24.

the Red Beds of the San Juan region and of the Anthracite, Crested Butte, and Tenmile quadrangles of Colorado. Alternation of paler with darker and with brighter shades of red is remarked in almost every occurrence of Red Beds. In all these instances it is significant that the color boundaries tend to follow bedding planes,

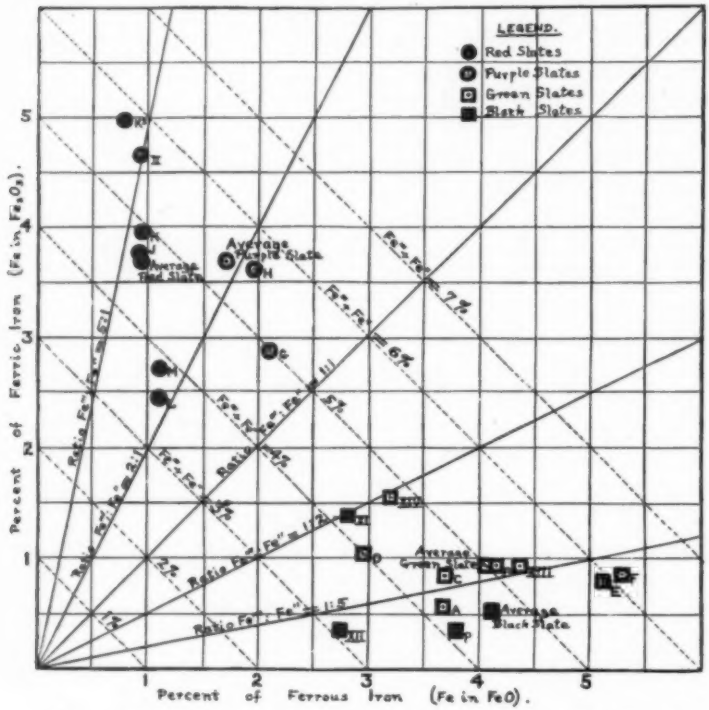

FIG. 2.—Diagram to illustrate the relation between color and the proportion of ferric to ferrous iron in ferruginous slates. Analyses (from Dale) are lettered to correspond with text.

and usually accompany changes in coarseness of grain. Many of the color boundaries are distinct, even planes, but they may be rendered irregular by downward migration of coloring matter.

Where alternations of light- and dark-red strata occur, the more deeply colored beds are in most cases of finer grain than the

others. The occurrence of coarse-grained massive buff sandstones in a series of maroon or chocolate shales has been noted by many writers. This association holds true in many other Red Beds besides the group here under consideration. Thwaites[1] reports it as an almost unfailing relation in the Lake Superior sandstone series of northern Wisconsin, and Geikie[2] mentions its existence in the Triassic New Red Sandstone of Great Britain. The respective tints are understood to be uniform throughout the strata in which they are noted, and not to be merely the surface staining of the beds, which might have a very different origin.

Distribution of greenish colors.—Greenish tints occur in the Red Beds as a fairly even color through continuous beds interstratified more or less closely with red strata, in streaks and blotches in red strata themselves, and occasionally in strips following joint planes. The completely greenish strata, and likewise the mottled beds, include both shales and sandstones, and are described from many districts. Green spots in red strata are as a rule irregularly ellipsoidal in shape, somewhat flattened parallel to the bedding, and indefinite in outline. They may vary in diameter from a few millimeters to several inches.

Variations in color along the strike are as common and as much to be expected as similar variations in texture, cross-bedding, or any other stratigraphic feature. The Red Beds of the Southwest are noted for their inconstancy and frequency of change along the strike; those of Central Wyoming and the Black Hills are fairly continuous in lithologic characters for considerable distances. Where other features are variable, the color is variable also; and where other features are constant, color likewise is constant.

Nature of the coloring matter.—Available data as to the chemical composition of the green bands and spots and other variations in color in the western Red Beds are very meager; but many of the same phenomena occur in roofing slates, whose commercial value has been the cause of painstaking investigation of their occurrence and character. The close dependence of color upon chemical

[1] F. T. Thwaites, "Sandstones of the Wisconsin Coast of Lake Superior," *Bull. Wis. Geol. and Nat. Hist. Survey No. 25*, 1912, p. 31.

[2] Archibald Geikie, *Text-Book of Geology* (London: Macmillan, 1903), p. 1064.

composition is brought out strikingly by the following analyses, which are at least sufficient to show that the differences in color in the Red Beds are caused by the same differences in chemical composition as those which cause corresponding differences in color in slates. The analyses are plotted graphically in the diagram (Fig. 2).

The following analyses are taken from Dale:[1]

		Fe_2O_3	FeO	Net Fe	Ratio Fe‴: Fe″	Free Carbon
Red slates	J	5.36	1.20	4.69	4.04 :1.00	
	K	5.61	1.24	4.89	4.09	(No free carbon found in any of the red or purple slates)
	L	3.48	1.42	3.54	2.29	
	M	3.86	1.44	3.83	2.42	
	K²*	7.10	1.00	5.75	6.37	
	Average	5.26	1.21	4.62	3.91 :1.00	
Purple slates	G	4.10	2.71	4.99	1.36 :1.00	
	H	5.16	2.54	5.59	1.84	
	II	6.63	1.20	5.58	5.00	
	Average	5.30	2.15	5.38	2.22 :1.00	
Green slates	A	0.81	4.71	4.23	0.156:1.00	
	B	1.34	5.34	5.09	0.226	Trace
	C	1.23	4.73	4.54	0.234	Trace
	E	1.12	6.58	5.90	0.155	
	F	1.24	6.81	6.16	0.165	
	O	1.47	3.81	3.99	0.348	
	XIII	1.33	5.64	5.31	0.213	Trace
	XIV	2.24	4.07	4.73	0.496	
	Average	1.35	5.21	4.99	0.232:1.00	
Black slates	P	0.52	4.87	4.15	0.095:1.00	0.46
	VII	.....	9.03	7.02	0.000	1.79
	XI	1.98	3.65	4.23	0.490	0.77
	XII	0.53	3.52	3.11	0.135	1.54
	Average	0.75	5.27	4.63	0.129:1.00	1.14
Spotted slates	M*	3.86	1.44	3.83	2.42 :1.00	
	Q*	1.79	1.19	2.18	1.34 :1.00	
	R*	1.09	1.06	1.58	0.927:1.00	

*K² is same slate as K, but finer grained. M = red slate near green spot. Q = purple rim of spot. R = green part of spot.

[1] T. N. Dale, "The Slate Belt of Eastern New York and Western Vermont," *Ann. Rept. U.S. Geol. Survey No. 19*, 1899, Part 3, pp. 232, 246–53, 257, 264; and "The Roofing Slates of the United States," *U.S. Geol. Survey Bull. No. 275*, 1905, pp. 34–36.

The following analyses are taken from W. J. Miller:[1]

VERNON FORMATION, CENTRAL NEW YORK

	Fe_2O_3	FeO	Net Fe	Ratio Fe‴ : Fe″
Red shale	2.25	0.75	2.16	2.72:1.00
Green spot in red shale	0.00	1.19	0.93	0.00:1.00

The following analyses were furnished by Eliot Blackwelder:[2]

CHUGWATER FORMATION, WIND RIVER VALLEY, WYOMING

	Fe_2O_3	FeO	Net Fe	Ratio Fe‴ : Fe″
Red sandstone	3.50	1.04	3.26	3.02:1.00
Greenish sandstone*	1.03	1.04	1.53	0.89:1.00

*The greenish sandstone of the Chugwater in this case is a strip of the ordinary red sandstone, leached along a joint crack.

The following analyses are taken from Richardson:[3]

SPEARFISH FORMATION, BLACK HILLS

	Fe_2O_3	FeO	Net Fe	Ratio Fe‴ : Fe″
Green shale	1.85	1.04	2.02	1.78:1.00
Red shale, adjacent to green	4.61	1.24	4.55	3.72:1.00
Red shale	3.64	0.65	3.05	5.04:1.00
Red shale	2.04	0.18	1.57	10.20:1.00

The black color in the slates is due, not directly to any peculiarity of the iron content, but to the presence of carbonaceous matter, which incidentally brings about the reduction of iron oxide to the ferrous form. Black shales are very rare in the western Red Beds, but highly carbonaceous and even coal-bearing strata occur in the Newark series of the Atlantic Piedmont. The typical Newark clastics are quite intensely red, many of them with a purplish tone, but the carbonaceous strata are always gray or black.

The color of the prevailing red strata in the Red Beds series is due to the presence of ferric oxide. The iron of the coloring matter

[1] W. J. Miller, "Origin of Color in the Vernon Shale," *Bull. N.Y. State Museum*, No. 140, in *63d Ann. Rept. N.Y. State Museum, 1909*, I, 150–56.

[2] U.S. Geol. Survey, Division of Chemistry, Analysis No. 2530.

[3] G. B. Richardson, "The Upper Red Beds of the Black Hills," *Jour. Geol.*, XI (1903), 365–93.

is chiefly in the ferric state in light- and dark-red, buff, and yellowish strata. A gray or green color, on the contrary, signifies a low proportion of ferric oxide, and usually a preponderance of ferr*ous* over ferr*ic* compounds. The mineral composition of the coloring matter is difficult to determine accurately because of its fine division. A green color is "common with silicates in which ferrous iron is prominent,"[1] and silicates may be important where the ratio of ferrous to ferric iron is high.

IS THE FERRUGINOUS MATERIAL AN ORIGINAL CONSTITUENT OF THE SEDIMENTS, OR A LATER INTRODUCTION?

Hypothesis of introduction of iron from igneous magmas.—If the ferruginous matter was an integral part of the original sediments, we have no more difficulty in explaining its presence than in accounting for any other common mineral constituent of sedimentary rocks. Its source was most probably in the same rocks which gave rise to other materials of the series, such as quartz, feldspar, and calcite, and the agencies of transportation were the same as those which were responsible for the entire series. If, however, we postulate that the iron has been introduced since the completion of sedimentation, after the series of Red Beds was otherwise complete, we shall find it very difficult to reconstruct in imagination any natural agency which might have brought about such a result, and we shall be puzzled to find an adequate source for this vast amount of iron.

There is one hypothesis of this kind which received credence and vigorous support in America for several decades of the nineteenth century, in explanation of the red stain in the Newark series of the Connecticut Valley.[2] The close association, in that series, of clastic red sediments with contemporaneous extrusive and intrusive basic igneous rocks of great thickness and extent made it a natural suggestion that the exceptional color of the sedimentary members was connected directly with igneous action. The absence of any

[1] E. S. Dana, *A Text-Book of Mineralogy* (New York: John Wiley & Sons, 1909), p. 472.

[2] See J. D. Dana, *Manual of Geology*, ed. 1880, p. 764; and also his review of Russell's bulletin on the "Subaerial Decay of Rocks" (*U.S. Geol. Survey Bull. No. 52*, 1889), in *Am. Jour. Sci.*, 3d Ser., XXXIX (1890), 319.

strong evidence in support of this hypothesis, aside from that of association with igneous rocks, and the discovery of various lines of evidence (among those outlined below) in direct opposition to it, however, caused it to fall into disrepute. In the case of the western Red Beds there is not even the association with igneous rocks to suggest such an explanation. There are no contemporaneous igneous rocks in any part of the Red Beds group, and in the greater part of the area in which the group occurs there are no later igneous rocks known. In no case is there a relation comparable to that in the Connecticut Newark.

Where the effect of later intrusions upon Red Beds has been observed carefully, that effect is not to heighten, but to destroy, the red color. This action has been noted in the Tenmile district[1] and in the Anthracite-Crested Butte district,[2] Colorado.

Later introduction of iron by meteoric waters.—There is no apparent reason why the Red Beds should have been favored by post-sedimentational iron-bearing solutions while other clastic series in the same region, both older and younger, were not stained. It might be expected that the coloring matter would be most abundant in the most pervious strata, especially along the lower surface of such strata, where they are in contact with less pervious rocks. It is quite true that color boundaries in the series follow bedding planes very closely; but, unfortunately for the hypothesis of later introduction, the more highly ferruginous strata of the Red Beds are as a rule, and with few exceptions, the more impervious. As noted above (pp. 158–59), in a series of alternating sandstones and shales it is almost invariably the shales which are deeper in color.

Ferric hydrates are not being introduced into the Red Beds strata along the present joint planes. The analyses of Chugwater sandstone given on p. 161 (see also footnote) show a marked leaching of iron along a joint. Nor is iron commonly concentrated along joints in paler sediments underlying the Red Beds.

[1] S. F. Emmons, Tenmile District Special Folio (No. 48), *Geol. Atlas of the U.S., U.S. Geol. Survey*, 1896.

[2] G. H. Eldridge, "Description of the Sedimentary Formations," Anthracite-Crested Butte Folio (No. 9), *Geol. Atlas of the U.S., U.S. Geol. Survey*, 1894.

Deposition of coloring matter contemporaneous with sedimentation. —If the ferruginous material which furnishes the color was concentrated in the sediments chiefly at the time of their deposition, this association of high color with fine sediment is explained readily, as follows:

In the weathering of igneous rocks, ferruginous material is separated out chiefly by the chemical decomposition of iron-bearing silicate minerals, and is therefore at the time of separation in a very fine state of division. At the same time ferrous salts of iron usually are altered to ferric oxide or hydrate. During surface transportation and sorting it is segregated in whole or in part, by reason of this fine division (if it persists) and in spite of its high specific gravity, along with other finely divided materials constituting muds and "clays."

This does not apply to iron occurring in the parent rock in the form of magnetite or other very stable minerals, which in most cases are concentrated with the coarser products of mechanical disintegration, such as sands and sandy shales. Ferruginous materials firmly cemented to sand grains during weathering may also be transported and deposited with the sand. Iron taken into solution will have yet a different history.

In the weathering of sedimentary rocks, the behavior of their ferruginous content is dependent upon the behavior of that material in a former sedimentary cycle, and thus ultimately upon the conditions already outlined for the weathering of the igneous rocks. Ferric oxide, the form in which iron occurs most abundantly in sediments, is, because of those conditions, usually finely divided, and in a second cycle of transportation and deposition will be concentrated again chiefly with the muds. In so far as assortment is imperfect, the ferruginous material may be deposited with any type of sediment.

Because of the usual absence of any commercial value in the series, there is an unfortunate dearth of analyses of Red Beds shales and sandstones. The fact of the concentration of ferruginous matter in the fine-grained sediments is well illustrated, however, by composite analyses of shales and sandstones

from all sources. The following figures from Clarke[1] are to the point:

	Percentage Fe_2O_3	Percentage FeO	Total
Average of 78 shales (mean of two composites) ..	4.03	2.46	6.49
Composite analysis of 253 sandstones.....	1.08	0.30	1.38

See also analyses K and K[2], on p. 160, *supra*.

Microscopic evidence.—Russell[2] has presented evidence to show that the red coloring matter in sandstones of the Newark series in Virginia occurs as a coating on the sand grains, and that it existed in that form even before the transportation of the sand from the residual soils from which it was derived. Since much of his argument would apply equally well to the red sandstones of western Red Beds, it is worth investigation in this connection. In a few thin sections of Newark sediments from Virginia which have been available for study by the writer, it is shown clearly that ferruginous matter exists there, both in the form described above and as a later interstitial filling between grains. Many of the sand grains, both of quartz and of feldspar, are well rounded. Most of them are surrounded by a thin coat of hematitic material, whose outer surface is smooth and even, but whose inner border may show slight irregularities penetrating into the body of the grain. This red film may not be entirely continuous around the grain; and this fact, together with the smoothness of its outer surface, suggests, as Russell maintained, that the coating was acquired by the sand grain before its final deposition, and that the coating suffered wear during transportation, without being completely removed.

In a number of instances it was found that part of the cementation of the rock had been accomplished by enlargement of the original grains of quartz and feldspar in optical continuity, outside of the red coating; and that outside of the enlargements there existed other bodies of hematitic matter, irregularly scattered

[1] F. W. Clarke, "The Data of Geochemistry," 2d ed., *U.S. Geol. Survey Bull. No. 491*, 1911, chaps. iii, v, vi.

[2] I. C. Russell, "Subaerial Decay of Rocks," *U.S. Geol. Survey Bull. No. 52*, 1889.

through the interstitial spaces. This material may have filtered in during the process of cementation; or it may have been present at the time of sedimentation and have been thrust out of the way by the growing crystals during enlargement. Occasional specks of red-stained material are found within the area of the enlargements and distinct from the inner coating of the original grains; and where the enlargement is missing, the interstitial hematitic matter may be difficult to distinguish from that of the primary coating. The stained areas probably do not represent pure ferric oxide, but ferric oxide in such ratio to silica or clayey material as to render the mixture nearly opaque and quite uniformly red or reddish brown; for chemical analyses of the same rocks show the percentage of ferric oxide in the rock to be much lower than the percentage of red-stained area in the sections.

A similar microscopic investigation has been made by Richardson[1] in his study of the Spearfish formation of the Black Hills. He says: "Amorphous red pigment is prominent in the slides. It irregularly coats and spots the minerals, and . . . . constitutes the chief interstitial substance."[2] He also presents an analysis to show that the ferric oxide of the pigment is essentially anhydrous.

Thin sections of Red Beds from all parts of the West are not available at present, and it has not been practicable, therefore, to make a thoroughgoing microscopic study of the group. Richardson's description suggests a relation of pigment to cementation similar to that found in the Newark sandstones: a twofold relation, indicating that part of the pigment was transported and deposited as a coating on sand-grains, and that the remainder constitutes an important part of the cementing material of the rock and is coeval with the rest of that material.

But what was the time of cementation? Was this process completed before the exposure of the Red Beds series to erosion, or is it still going on, or did it cease at some intermediate time? In so far as this question affects the iron content of the Red Beds we really have answered it already; for if any considerable part of the iron had been introduced as a cement later than the time of sedimentation it would not be found most abundantly in the strata which

[1] *Op. cit.* [2] *Ibid.*, p. 379.

afford the least ready passage to water, as is actually the case. The cementing material of the Red Beds, in so far at least as it is ferruginous—and the remainder of it has little or no bearing upon the point in question—was therefore for the most part present in those beds at the time they were deposited as sediments.

Extent of known secondary redistribution of coloring matter.—Although the evidence is convincing that the ferruginous matter in the Red Beds was an integral part of the original sediment, and that it was deposited originally in the series in practically its present distribution and arrangement, it is equally certain that there have been some later modifications of that primary distribution. A large majority of all variations of color within the Red Beds are limited by bedding planes; but there are numerous minor variations in color due to the migration of coloring matter along the lines of movement of ground-water.[1] Reducing solutions locally extract ferruginous matter along joints, or cause a general downward movement of iron. In the western Red Beds, as in Barrell's section of the Catskill formation in Schuylkill County, Pennsylvania,[2] the lower boundaries of many deeply stained bands are drawn less sharply than the upper. All of these are very minor features, however, in the distribution of the coloring matter of the series as a whole. The analyses (p. 161) furnished by Blackwelder illustrate the effect of leaching along a joint; the color changes from red to greenish, the net iron content is reduced from 3.26 per cent to 1.53 per cent, and the ratio of ferric to ferrous iron drops from 3.02:1.00 to 0.89:1.00.

IS THE FERRUGINOUS MATERIAL IN THE SAME FORM AS AT THE TIME OF SEDIMENTATION?

Variations in degree of oxidation.—Is the color due to recent oxidation in weathering? It has sometimes been asserted that the color of the Red Beds is a superficial matter, due to the weathering of originally dull-colored sediments. In support of this idea drill records have been quoted, showing that beyond a certain depth

[1] Cf. Richardson, *op. cit.*, p. 376.

[2] See Joseph Barrell, "The Upper Devonian Delta of the Appalachian Geosyncline," *Am. Jour. Sci.*, 4th Ser., XXXVI (1913), 437 ff.

below the surface the rocks are no longer red. S. F. Emmons states, respecting the Maroon formation of the Tenmile district, Colorado,[1] that "in depth, as shown in underground workings, the red color generally gives way to a greenish gray." In another paragraph of the same folio, Emmons states that igneous or hot-water alteration destroys the red color. Inasmuch as the ores of this district are related intimately to igneous and hot-water action, it is natural to suppose that mine workings would be the most likely of all places in which to find such effects. The relation here described is probably a local and abnormal phenomenon.

In his description of the sedimentary rocks of the Anthracite and Crested Butte quadrangles, where the general situation is similar to that in the near-by Tenmile district, Eldridge[2] states that "the upper division [of the Maroon conglomerate] is of a peculiar red or chocolate color, except in regions of local metamorphism" of the kind mentioned also by Emmons.

Drilling explorations in the oil regions of Oklahoma and Texas recently have given us some additional information on the behavior of the color with depth. Gould[3] tells of a well some 3,300 feet deep in southeastern Oklahoma, in which "the last of the typical Red Beds was encountered" at a depth greater than 2,000 feet. The underlying strata were dull-colored sediments like those outcropping along the eastern (lower) margin of the Red Beds. Nowhere does he mention any change in color due to depth.

A well sunk 3,095 feet at Ashland, Wisconsin, passed through typical red sandstones of the Lake Superior group all the way, without any suggestion of a change in color with depth.[4]

It is to be remembered that most of the western Red Beds series are not colored uniformly throughout, but include many lighter-colored strata; and that many of the Red Beds successions include gray and green members. Any drill cutting through such a series would find, of course, changes in color with depth, but they would not be progressive, and they would have no causal connection with depth whatsoever.

[1] *Op. cit.* [2] *Op. cit.*

[3] C. N. Gould, "Petroleum in the Red Beds," *Economic Geology*, VIII (1913), 768–80.

[4] Data from F. T. Thwaites, *op. cit.*

This variegation itself constitutes one of the most unanswerable arguments against weathering in the present cycle as the cause of the red color. The distribution of gray and green beds among the red does not appear to have any definite relation to coarseness of grain, as in the case of intensity of the red color, nor to any other single stratigraphic feature; and it certainly bears no constant relation to the topographic surface.

Origin of mottling.—The green spots present in some otherwise red strata have been explained, on the hypothesis that the red color is due to recent oxidation in weathering, as remnants of the original color of the beds; but, as will be shown presently, it is much better in accordance with fact to explain them as spots in originally red sediments, deoxidized by the agency of some particle of organic matter which was present in the original sediment.

Dale[1] discusses the origin of the green spots in red and purple slates in part as follows:

> The difference in color from the green to purple to red is manifestly due to the differences in the amount of hematite. [See analyses, p. 160.] The green fossil impressions in purple slate may throw some light on the origin of these spots. In this case the effect of organic matter, whether the carbonaceous matter of the lining of an annelid boring or from a marine alga, has been to diminish the quantity of Fe_2O_3 in the slate. The increase of the carbonates may be directly connected with the production of CO_2 by decaying organisms and the consequent decrease of the Fe_2O_3. In view of all these facts and indications, the spots may be safely regarded as probably produced by chemical changes consequent upon the decay of organisms.

The same conclusion is reached by Miller with respect to green spots in the red Vernon shale (Silurian, central New York). Miller[2] finds dark organic centers in many of the spots, and attributes the color of green shales associated with the red strata to more abundant dissemination of organic matter.

It is probable that a very small quantity of organic matter may reduce or prevent the oxidation of a considerable amount of iron. The ferruginous matter necessary to stain a sediment is so small in amount that the quantity of organic matter necessary to accomplish the reduction of a patch less than an inch in diameter, like

[1] T. N. Dale, The Slate Belt of Eastern New York and Western Vermont, *Ann. Rept. U.S. Geol. Survey No. 19*, 1899, Part 3.

[2] *Op. cit.*

most of the green spots in Red Beds, is so small that it could easily be effaced or removed. A tiny fragment of vegetable fiber or the remains of a few minute organisms of any kind probably would suffice.

Cause of gray and green bands in red beds: Barrell's hypothesis.—Barrell[1] states that in the Catskill formation of eastern Pennsylvania gray and green colors are typical of sandstones, and red colors of shales. He therefore suggests a causal relation between coarseness of grain and condition of the iron content, as follows:

> These relations show that there was a tendency toward deoxidation during the formation of the beds of sand, of oxidation during the deposition of the Catskill muds. Where the clay and iron oxide were sparing in quantity, the deoxidation was effective. The conditions which accompanied the deposition of clay and iron oxide also permitted oxidation to dominate over deoxidation.[2]
>
> The lack of oxidation of the iron in the sandstones, in spite of its lesser quantity, suggests that more abundant ground-waters in the sands may have kept out the air and permitted the organic matter to accomplish its effects, or perhaps that here the ratio of organic matter was in excess of the ferric oxide.[3]
>
> A few rare carbonaceous streaks have been observed in the Catskill and the plant impressions are in places found in deoxidized shales. Coaly and pyritiferous plant fossils are also preserved in some of the olive sandstones.[4]

No actual remnants of organic matter are reported to have been found in *red* strata, though markings interpreted by Barrell as rootlet marks are noted by him in certain horizons of red shales.

That physical conditions of deposition alone should have favored oxidation in the finer-grained sediments and retarded it in the coarser seems highly improbable. Where meteoric water moves most freely, there the most oxygen is carried in solution. The more rapid the circulation of ground-waters, the more effective are those waters as oxidizing agents, instead of vice versa, as suggested by Barrell.

For the lesser quantity of iron in the sandstones we already have offered an explanatory hypothesis (p. 164). The smaller the amount of ferric oxide present in a sediment, the smaller, of course, is the amount of organic matter necessary to bring about its reduction—or the reduction of a sufficient part of it to mask the color of the remainder. If organic matter were distributed equally among

[1] *Op. cit.* [2] *Ibid.*, p. 458. [3] *Ibid.*, p. 469. [4] *Ibid.*

all the sediments of the series, it would cause gray and green colors in the sands rather than in the muds, by reason of the mechancial concentration of ferric oxide in the latter. The ratio of ferruginous materials to organic matter seems to be the dominating factor in determining the colors of ferruginous beds. Barrell's second suggestion in the foregoing quotation, dealing with this point, deserves more careful investigation.

Much of the organic matter carried as sediment in surface waters tends to be concentrated with the muds, for reasons analogous to those already suggested in the case of ferruginous material (i.e., fine division); but the proportion so concentrated is probably smaller in the case of organic matter than in that of ferruginous material, because of the indiscriminate distribution of driftwood, the growth of vegetation in most regions of clastic deposition, and other similar factors. In any of these cases organic matter in large quantities may enter into the composition of sands and muds alike, and thus bring about a higher ratio of organic to ferruginous matter in the sands than in the muds.

These controlling factors are so complex that no constant relation between coarseness of grain and distribution of organic matter is to be expected. It is evident from the quotations on p. 170 that definite organic remains in the section there under consideration are confined chiefly to gray and green strata; and it is also evident that such strata include both shales and sandstones. An examination of the detailed section[1] of the Catskill formation published by Barrell in this paper reveals a somewhat less uniform relation between coarseness of grain and color of beds than one would understand to be the case from reading the text or the labels on the graphic columnar section.[2] Of the 3,864 feet of beds definitely described as of one or the other type, 2,179 feet (56.4 per cent) bear out Barrell's generalization that gray and green colors are typical of sandstones and red colors of shales, 901 feet (23.3 per cent) are noncommittal, and 754 feet (20.3 per cent) are in opposition to the rule. This variability is in better accord with the complexity of the factors controlling the distribution of organic matter in sediments than a more constant relation would be; and suggests

[1] Barrell, *op. cit.*, pp. 451–56. [2] *Ibid.*, p. 457.

that even here the distribution of organic matter may be a controlling influence in determining the colors of the individual strata.

Dawson,[1] in discussing the Triassic (?) Red Beds of Nova Scotia, says of the gray sandstones and shales interstratified with them, that "where thick, they always contain either fossil plants, bituminous matter or thin seams of coal, or all of these. The following sentence from Geikie,[2] relative to the Old Red Standsone of the British Isles, is also interesting in this connection: "It may be observed also that where gray shales occur intercalated among the red sandstones and conglomerates they are often full of plant remains, and may contain also ichthyolites and other fossils which are usually absent from the coarser red sediments."

Organic matter the controlling influence in the case of the western Red Beds.—Nowhere in the literature on the western Red Beds is there suggested such a definite and relatively constant association of green and gray colors with sandstones, and of red with shales, as that which Barrell sees in the Catskill formation, and as that which is described as occurring in the Siwalik formation of India.[3] In the foregoing quotation from Geikie, the opposite relation is implied in the Old Red Sandstone series of Great Britain. In the Red Beds of the western United States variegation is perhaps more common in shales than in sandstones, though it occurs to a marked extent in both.[4] The distribution of gray and green colors in the Red Beds coincides very closely with the distribution of organic remains in the same series, in so far as such remains are present; and this close association, together with the chemical probabilities of the case, suggest that organic remains now obliterated explain at least the greater part of the remaining gray and green areas and strata. The decolorization of these sediments may, therefore, have been complete before their burial under later strata. The

[1] J. W. Dawson, "On the Colouring Matter of Red Sandstones and of Greyish and White Beds Associated with Them," *Quar. Jour. Geol. Soc. London*, V (1848), 25–30. Quotation from p. 26.

[2] Geikie, *op. cit.*, p. 1003.

[3] Medlicott and Blanford, *A Manual of the Geology of India*, II (1879), 524–26. Quoted by Barrell, *op. cit.*, pp. 463–64.

[4] Cf. Permian of the Pecos Valley and of the Zuni and Colorado plateaus; and the Jura-Trias Painted Desert sandstone of the latter plateau.

occurrence of traces of organic matter in red strata may be explained by an unusually high content of ferric oxide in those strata, or by later reoxidation of iron at one time in the ferrous state. The fact that much the greater part of the occurrences of colors characteristic of ferrous compounds are in distinct beds with definite boundaries, indistinguishable in most other characteristics from other beds which are not so colored, is good evidence that this distribution of ferrous and ferric compounds, or of the substances responsible for these compounds, was accomplished for the most part at the time of sedimentation. It is well to remember that the gray and greenish strata are very subordinate in the Red Beds of the western United States.

The general conclusion to be drawn from the preceding discussion is that there has been in the western Red Beds no extensive change of ferrous to ferric iron, or vice versa, since the time of sedimentation; and also that the original distribution of these compounds in the series was influenced most largely by the distribution of organic matter.

Variations in hydration of ferric oxide.—That various degrees of hydration exist in the ferric oxide of the Red Beds today is clear from the variety of red, brown, and yellow hues which appear in some members of the group. The major part of the ferric oxide in the Red Beds is no doubt but poorly hydrated.[1] The bright- and deep-red and red-brown colors (which are most common in the western series) may be attributed to hematite (anhydrous Fe_2O_3) or to turgite ($2Fe_2O_3 \cdot H_2O$). The lighter browns, yellow-browns, and yellow tints are referable to göthite ($Fe_2O_3 \cdot H_2O$) or limonite ($2Fe_2O_3 \cdot 3H_2O$), or possibly in some cases even to xanthosiderite ($Fe_2O_3 \cdot 2H_2O$).

The freer passage of water through the sandstones as compared with the shales makes the constituents of the former, after consolidation, more susceptible of hydration than those of the latter. It is entirely probable that in some cases at least this factor of porosity heightens the contrast in color between coarse and fine sediments; but it apparently has not affected the greater part of the Red Beds, in which the ferric oxide is relatively anhydrous. Van

[1] Cf. Richardson's investigation, discussed on p. 166.

Bemmelen has shown that chemically prepared ferric hydrate, after being partially dehydrated, if placed in a medium saturated with water vapor, at ordinary temperature, takes up again part of the lost water.[1] Brescius went so far as to say that "when nearly dry, ferric hydrate has almost as great a tendency to take up water as oil of vitriol itself."[2] This does not appear to be true, however, of hematite found in nature. Once *completely* dehydrated, ferric oxide becomes a stable compound.

It remains to inquire into the means by which dehydration of the more hydrous compounds might have been accomplished to produce the low hydrates, in case these were not originally in the same condition. The first agent of dehydration which presents itself is that of heat. Elsden makes the following statement:

> The influence of temperature and moisture upon the iron hydrates is well known. In the case of laterite in India, the yellow xanthosiderite soon weathers to reddish-brown turgite, owing to dehydration. In the hot and arid regions of South California the soils are dark red in colour, the iron being in the form of hematite instead of the hydrous forms, göthite or limonite. Dehydration also takes place in the hot regions of the Southern Appalachians, where the air is comparatively humid. It is only the deeper portions of the soil which retain the iron in a hydrated form.[3]

In all of the above mentioned cases, the source of the heat which produces the reaction in question is the sun's rays. Its action in the soils is limited to a superficial stratum rarely more than 15 feet in maximum depth.[4] Obviously, the direct influence of insolation cannot be responsible for any extensive dehydration in the Red Beds, whose characteristic colors are known to extend to depths of more than 2,000 feet.[5]

[1] J. M. Van Bemmelen, "Sur le colloide de l'oxyde ferrique," *Recueil des travaux chimiques des Pays-Bas et de la Belgique*, VII (1888), 112.

[2] E. Brescius, "Researches on Ferric hydrate," abstract in *Jour. Chem. Soc. London*, XXIV (1871), 497.

[3] J. V. Elsden, *Principles of Chemical Geology* (London: Whittaker & Co., 1910), p. 97.

[4] W. O. Crosby, "On the Contrast in Color of the Soils of High and Low Latitudes," *Amer. Geologist*, VIII (1891), 74; E. A. Smith, *Geol. Survey of Alabama, Rept. for the Years 1881 and 1882*, p. 186.

[5] See p. 168.

Another source of heat which might be suggested is that of igneous intrusions. The absence of igneous rocks in the greater part of the western Red Beds, including the regions where drilling has shown the colors to be unchanged at depth, proves that this has not been a factor of widespread importance. Heat due to regional metamorphism or to structural deformation of any sort must likewise be discredited here, as the Red Beds are substantially flat-lying over vast areas, and are nowhere intensely deformed or metamorphosed, except locally in the immediate neighborhood of igneous bodies.[1] That compression due to the weight of overlying sediments may have created sufficient heat for the accomplishment of extensive dehydration must be recognized as a possibility, although in many areas where the Red Beds occur the requisite overlying sediments are not known ever to have existed, and the uppermost members of the series in these localities are as brilliantly red as any below them. Furthermore, pressure creates heat only by the performance of mechanical work, and microscopic study of the Red Beds reveals no evidence of internal deformation.

Crosby[2] concludes that the process of dehydration of ferric oxide is largely a spontaneous one, which goes on independently of any outside influence whatsoever, though aided by high temperatures. He states as evidence in support of this hypothesis that the red sedimentary formations and the red iron ores of the world occur in the older systems chiefly, while in the younger systems ferruginous formations and ores are commonly yellow. There are, however, many exceptions to this rule, such as the buff Cambrian sandstones of the Mississippi Valley, the modern red residual iron ores of Cuba, and the dark-red hematitic bog ores in Sweden and elsewhere; and furthermore, the dehydration of the pre-Cambrian red sandstones and argillites may be attributed in some cases partly to regional metamorphism, which has not affected the younger beds.

Richardson[3] has given much weight to this "essentially spontaneous" process. He states that "the dehydration of ferric hydrates tends to go on under ordinary conditions without any unusual cause."[4] "This has been repeatedly demonstrated by experiment."[5] But none of the experimenters to whom Richardson

[1] See p. 168. [2] *Op. cit.* [3] *Op. cit.*, p. 392. [4] *Ibid.* [5] *Ibid.*

refers[1] appears to have considered the observed dehydration as truly "spontaneous"; and certainly they have shown the process to be closely dependent upon external conditions. The chemically precipitated colloidal ferric hydrates possess, when first formed, a higher content of water than any of the forms known to occur as minerals. According to Van Bemmelen,[2] on standing in a dry medium at ordinary temperatures these colloids gradually approach the composition $2Fe_2O_3 \cdot H_2O$, beyond which the percentage of combined water is not reduced without the application of much higher temperatures. Under water, or in air of moderate humidity, they have not been shown to lose water beyond the composition $Fe_2O_3 \cdot 2H_2O$ except at temperatures above 50° C.,[3] and heating at 50–60° C. for 2,000 hours failed to bring about dehydration beyond the composition $2Fe_2O_3 \cdot H_2O$.[4] Temperatures as high as these cannot be assumed to have existed, except locally, in the Red Beds sediments since burial. The so-called spontaneous dehydration observed in the laboratory is probably subject to the terms of Van Bemmelen's conclusion: "The red-brown substance, which has been considered to be a hydrate, is a colloid . . . . which has no stable composition; it maintains an equilibrium with the tension of the water vapour in the surrounding medium."[5] Since burial, the great mass of the Red Beds sediments, except in the most arid districts of the West, have been saturated with ground-water, a condition decidedly unfavorable to dehydration at the temperatures there existing.

Yet another agent of dehydration is mentioned by Elsden:

> The presence of any substance in solution which lowers the vapour tension of water will lower the inversion temperature of gypsum. . . . . Even solid

[1] J. M. Van Bemmelen, "Sur le colloide de l'oxyde ferrique," *Recueil des travaux chimiques des Pays-Bas et de la Belgique*, VII (1888),106–14; Edward Davies, "Action of Heat on Ferric Hydrate in Presence of Water," *Jour. Chem. Soc. London*, XIX (1866), 69–72; G. C. Wittstein, "Über das Verhalten des Eisenoxyhydrates unter Wasser," *Vierteljahreschrift für praktische Pharmacie*, I. Band (1852), 275–76.

[2] *Op. cit.*, pp. 110–11.

[3] T. Carnelly and James Walker, "The Dehydration of Metallic Hydroxides by Heat," *Jour. Chem. Soc. London*, LIII (1888), 89; D. Tommasi, "Ferric Hydrates," abstract in *Jour. Chem. Soc. London*, XLIV (1883), 24.

[4] Davies, *op. cit.*, p. 70.

[5] Van Bemmelen, *op. cit.*, p. 114. Translated from the French.

gypsum . . . . can be changed into anhydrite by a concentrated solution of sodium chloride. . . . . These facts are of interest as pointing to the possibility of dehydration of minerals in rocks, in contact with salt solutions, at a temperature considerably below their normal inversion point.[1]

. . . . The occurrence of red ferruginous sandstone in conjunction with layers containing brown hydrated ferric oxide is less readily explained [than the superficial dehydration of soils, mentioned in the quotation on p. 174, *supra*], but the dehydration of certain beds may have been effected by contact with salt solutions, as in the case of gypsum already referred to above.[2]

Dehydration by contact with salt solutions presumably would affect the more porous beds first; whereas in the western Red Beds, as already described,[3] it is generally in the more porous beds that the lighter colors occur, and the shales are in general of deeper hue than the sandstones. Just how far the action of salt solutions may have been effective, both during and since sedimentation, in accomplishing dehydration in the neighborhood of such saline deposits as occur in the Red Beds, it would be difficult to say; but for the group as a whole it appears that this agency cannot have been of general importance.

In summation, it may be said that while widespread dehydration of iron oxide in the Red Beds since sedimentation cannot, at present, be proved not to have taken place, the greater weight of evidence now at hand is opposed to it; that the opposite process, hydration, may well have been active in the more pervious beds of the series; and that, therefore, the probabilities are quite as much in favor of a lower degree as of a higher degree of hydration, on the average, in the Red Beds at the time of sedimentation than at present.

WAS THE COLORING MATTER A CHEMICAL OR A MECHANICAL SEDIMENT?

Having determined as nearly as the available evidence permits the condition of the coloring matter of the Red Beds at the time of their deposition, we may proceed to inquire as to the geographic conditions which gave rise to sediments so colored. The first

[1] Elsden, *op. cit.*, pp. 85–86. See also H. Stremme, "Zur Kenntnis der wasserhaltigen und wasserfreien Eisenoxydbildungen in den Sedimentgesteinen," *Zeit. für prakt. Geol.*, January, 1910, pp. 18–23.

[2] Elsden, *op. cit.*, p. 97.

[3] P. 158–59.

question to be answered in this connection is: Was the ferruginous matter deposited as a mechanical or as a chemical sediment?

The general absence of coloring matter from the non-clastic members,[1] which are in very large part inorganic chemical precipitates, indicates quite clearly that the conditions favoring free chemical deposition of calcium and magnesium carbonate or of calcium sulphate were not those under which the coloring matter was usually deposited. The close limitation of ferruginous material to the clastic sediments proves that the conditions under which clastic sedimentation took place favored the deposition of iron oxide also, and strongly suggests that that material itself was carried and deposited as a mechanical sediment, for the most part at least. In this connection it is of interest to note that Dawson[2] drew a similar conclusion as early as 1848 with reference to the contrast in color between the clastic and non-clastic strata of the Red Beds of Nova Scotia.

The condition of the ferric oxide in the Red Beds sediments, as revealed by the microscope, is one of very fine division. Since fine division is to be expected from the mode of origin of the ferric oxide in soils,[3] this cannot be taken as evidence that it is a chemical precipitate, in place, in the rocks. If it were the latter, definite orientation of crystals of hematite with respect to peripheries of grains might be looked for. I never have seen this phenomenon in thin sections of Red Beds sediments. The microscopic evidence is therefore rather noncommittal as regards the present question.

The processes of weathering leave much the greater part of the iron content of all types of rock as a residue in the soil, subject to mechanical transportation. All of the other common chemical constituents of rock, with the probable exception of alumina, are leached from the soil more rapidly than ferric iron. The scarcity of iron in any form in the surface waters of the continents, abundantly shown by the analyses of lake and river water published by Clarke,[4] and its even greater scarcity in the ocean,[5] testify to the

[1] See p. 157. [2] *Op. cit.* [3] See p. 164.

[4] F. W. Clarke, "The Data of Geochemistry," 2d ed., *Bull. U.S. Geol. Survey No. 491*, 1911, chap. iii.

[5] *Ibid.*, chap. iv.

fact that chemical deposition of salts of iron is only exceptionally an important process in earth metamorphism. Glauconite, the most common of such deposits at the present time, is of such peculiar nature as to be readily recognized where it occurs in older sediments: it certainly is not involved to any appreciable extent in the origin of the Red Beds. Bog iron ore, the only other common type of ferruginous chemical precipitate at present, is connected intimately, in origin, with abundance of vegetation and with peculiar and limited topographic conditions; and although deposits of this type are scattered over many parts of the world, none of them is comparable in extent or in thickness to even the smallest of the Red Beds areas. Furthermore, no deposits similar in textural character to bog ores are known in the western Red Beds.

In view of the facts above stated, it seems a safe conclusion that the coloring matter of the Red Beds was transported and deposited almost if not quite wholly as a mechanical sediment; and, therefore, without danger of serious error, we may limit investigations of possible conditions of origin of this coloring matter to those which would produce it as a mechanical sediment purely. This applies to the gray and green members of the Red Beds series as well as to those in which the iron is present chiefly as ferric oxide; for if the ferrous iron in the former be explained as the result of the action of organic matter deposited in those strata,[1] the ferruginous matter may have been in the ferric form during transportation, quite as well as in any other.

[1] See pp. 170–72.

[*To be continued*]

STUDIES IN HYDROTHERMAL ALTERATION

PART I. THE ACTION OF CERTAIN ALKALINE SOLUTIONS ON FELDSPARS AND HORNBLENDE

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In the study of ore deposits from a genetic standpoint the subject of attendant wall-rock alteration has received deserved attention from geologists. Profound changes of a chemical and mineralogical character have been recorded at many places, and the relation existing between the various types of wall-rock alteration accompanying ore deposition led to the suggestion that a knowledge of the conditions which bring about such alteration would throw great light on the problem of ore genesis. The data obtained in regard to the temperatures, pressures, and nature of solutions would also be of value in interpreting the geologic history of such occurrences.

The most important of these alteration minerals are kaolin, sericite, and chlorite, and the knowledge concerning their origin is chiefly confined at present to speculations based on their modes of occurrence and their associated minerals. There is especially great difference of opinion[1] as to the origin of kaolin. By some writers kaolinitic alteration is attributed to the action of meteoric waters rich in carbonic acid, by others to meteoric waters which have made a cycle of underground courses, and by still others to the emanations from a cooling magma while these are possibly yet in a gaseous state. In regard to sericite and chlorite associated with ore bodies there is nearly general agreement that these have been formed by the action of the solutions which deposited the primary minerals, whether these solutions be magmatic or meteoric, upon the feld-

[1] Bibliographies of the literature on kaolin are given by Rösler, *Neues Jahrb.*, Beil. Bd. XV (1902), 231, and by Lazarevic, *Zeit. prakt. Geol.*, XXI (1913), 345. An introduction to a discussion of the origin of kaolin was initiated by Lindgren, *Econ. Geol.*, January, 1915.

spars and the ferromagnesian minerals. Calcite and quartz are also prominent products of these reactions. Further, the relations of these minerals to the ores indicate that the processes of alteration and ore deposition have gone on contemporaneously. This is shown by the progressive decrease in the intensity of the alteration laterally from the veins, and by changes in the character of the alteration in the same direction. For example, the following extract from the report by F. L. Ransome[1] on the "Economic Geology of the Silverton Quadrangle" illustrates these points.

At points 150 feet east of the lode the country rock is fine-grained and faintly mottled, showing only a few pale phenocrysts of feldspar and an occasional tiny grain of quartz. Under the microscope the rock reveals the character of a much-altered andesitic tuff or fine breccia. The feldspars have been completely altered to aggregates of sericite and calcite, while areas of calcite and chlorite probably represent former phenocrysts of augite. The groundmass is a rather indistinct aggregate of secondary quartz, sericite, and chlorite with a little apatite and rutile. The rock is wholly recrystallized into a secondary aggregate while retaining the gross structure of the original.

At a distance of 100 feet from the vein the . . . . chlorite and calcite are abundant, but much of the plagioclase is still recognizable. Sericite and quartz are not such prominent constituents. . . . .

At 50 feet from the vein . . . . the feldspar phenocrysts have been changed to aggregates of calcite and sericite, while areas of chlorite and calcite with sometimes rutile are all that remain of the phenocrysts of augite or biotite. The groundmass also, while preserving the outlines and in small part the substance of former lath-shaped feldspars, is now an aggregate consisting chiefly of quartz, chlorite, sericite, and a little rutile and apatite. . . . .

At 2 feet from the vein . . . . it is seen that alteration has been more thorough. . . . . The forms of the phenocrysts are preserved by pseudomorphous aggregates of sericite with some chlorite, calcite, and rutile apparently after biotite, and quartz, sericite, and chlorite in varying proportions after augite and plagioclase. The groundmass is entirely recrystallized . . . . and the dominant minerals are quartz and sericite.

A specimen taken from the wall of the vein showed more evident alteration . . . . and the rock is wholly recrystallized. The former phenocrysts of feldspar are replaced by pseudomorphus aggregates of quartz and sericite. . . . . Of the augite no trace remains, but some sericite inclosing rutile is apparently pseudomorphous after biotite. The groundmass is a finely crystalline mosaic of quartz and sericite. The notable feature of this wall rock is the absence of calcite and chlorite. . . . .

[1] F. L. Ransome, *Bull. 182, U.S. Geol. Survey*, pp. 116–18.

To sum up then, the alteration involves the change of the feldspars to sericite, calcite, and quartz; of augite to calcite and chlorite; and of biotite to chlorite, sericite, and rutile.

Chloritic alteration appears to precede sericitic alteration and to require less intense or less prolonged action or solutions of a different character, so that sericite is found closer to the veins than chlorite, and chlorite dominates farther from the veins. It may be in place here to note that many consider the solutions which emanate from the magma in its final stages acidic in character while others hold that they are alkaline. A study of volcanic gases and of the sublimates present in craters indicates that the volcanic vapors are quite certainly acid, and Day and Shepherd[1] found that the magmatic waters which they collected from the small dome within the crater at Kilauea were acid. However, to conceive of these solutions as remaining acid for a long journey through rock masses after their escape from the magma requires a high degree of acidity, and geologists have been loath to accept such a view.

It is clear then that the experimental formation of these alteration products from the feldspars and ferromagnesian minerals will give some clew as to the nature of the solutions and the temperature and pressure conditions that obtained during the deposition of the associated primary ore bodies. A study of the literature bearing on the subject of the hydrothermal alteration and syntheses of various silicates leaves the reader somewhat at a loss to discern the geologic application of much of the data. Some of the experimental work is of great interest chemically but not geologically. Hydrothermal investigations that begin with mixtures of oxides instead of with distinct mineral species may have in many cases comparatively little geologic significance. When the definite mineral can be prepared synthetically from pure materials, nearly ideal conditions may be obtained, with results that are beyond question. However, since many of the minerals whose investigation is of paramount interest and importance in their relation to the analysis of geologic processes cannot be prepared synthetically, the next

[1] A. L. Day and E. S. Shepherd, *Bull. Geol. Soc. Am.*, XXIV, 593.

[2] This literature is ably reviewed by G. W. Morey and Paul Niggli in *Jour. Am. Chem. Soc.*, XXXV, No. 9 (September, 1913), 1106–30.

best course to adopt is to use minerals of as high degree of purity as are available. The fusion of such a mineral before its investigation must be vigorously condemned because it is well established that the minerals which crystallize from a melt are not always like those which were originally fused, and in many cases several distinct minerals are produced by this process from a single species. If the glass that is formed by the fusion is used, it may differ chemically from the original mineral by the loss of volatile gases, and it must differ in its energy content.

For example, Lemberg's[1] important work is in part open to these criticisms, though the volume of data supplied by him is remarkable. Other work of this sort has been carried on in glass containers which of themselves may have furnished a most important source of error. Further, the concentrations of the solutions which have been used, as a rule, have been far in excess of those known to exist in ground or hot spring waters. Just how much difference this would make in the results obtained is not yet clear, though it is highly probable that the solutions as they come direct from the magma are far more concentrated than those which appear at the surface in the form of hot springs. It is also recognized that such solutions, after having been in contact with the wall rock throughout their courses, have probably changed notably their compositions and concentrations and may bear but small traces of some of their original constituents. As a result of these considerations it is at present an open question how much geologic significance is to be attached to investigations carried on with such concentrations as Lemberg used, though his results have thrown much light on methods of attacking the problem of mineral alteration. Other investigators, especially Thugutt,[2] Königsberger and Müller,[3] Friedel and Grandjean,[4] Baur,[5] and Chroustschoff,[6] have contributed highly suggestive data.

[1] J. Lemberg, *Zeit. deut. geo.. Ges.*, Vols. XXXV, XXXVII, XXXIX, LX.

[2] J. S. Thugutt, *Zeit. anorg. Chem.*, II, 64–107, 113–16; *Neues Jahrb. Min. Geol.*, Beil. Bd. IX, 554–624.

[3] Königsberger and Müller, *Centralbl. Min.*, 1906, pp. 339–48, 353–72.

[4] Friedel and Grandjean, *Bull. Soc. Min.*, XXXII, 150–54.

[5] Emil Baur, *Zeit. physik. Chem.*, LXI, 567–76; *Zeit. anorg. Chem.*, LXXII, 119–61.

[6] K. von Chroustschoff, *Compt. Rend.*, CXII, 677–79.

At the suggestion of Professor W. H. Emmons, the writer began in 1912 a series of investigations of a hydrothermal character with particular reference to the geologic application of the results. It was hoped to contribute something concerning the temperatures, pressures, nature, and concentrations of the solutions which produce new materials from the feldspars and from some of the ferromagnesian minerals.

PLAN OF THE WORK

The plan of the work has been to try the effect of various simple dilute solutions upon the feldspars and one ferromagnesian mineral at various temperatures and pressures. This has been done up to 280° C., at intervals of about 50°. It is planned to work on up to about 500° C. if the results justify such a course. The solutions have been tried at various concentrations, for different lengths of time, besides the different temperatures.

The minerals used have been nearly pure species. A high-grade adularia from the St. Gotthard tunnel, albite from the Amelia Court House locality, orthoclase from Elam, Delaware County, Pennsylvania, microcline from C. A. F. Kahlbaum (locality not known), and an alumihous hornblende from Renfrew, Ontario, Canada. These have in all cases been powdered to pass a 100-mesh sieve, and portions exactly or approximately one gram in weight have been used for each experiment.

THE OVEN

For work at temperatures above 100° C., the oven described as follows was constructed (Fig. 1). A box $28\frac{1}{2}\times15\times16$ inches (outside measurements), made of sheet iron, is fastened at the edges to a similar box $26\frac{1}{2}\times13\times15$ inches placed inside of it, with the space between packed with asbestos. A door at the front extends the full length and height of the oven, opening outward with hinges along the bottom. Two holes are cut in the top at H and H', Fig. 2, for the insertion of thermometers. A number of holes, o and o', about $1\frac{1}{2}$ inches in diameter are also cut in one end of the box, Fig. 3. When the oven is in use these holes are closed with loosely packed asbestos wool.

The heating arrangement consists of six coils of No. 22 nichrome wire wound on asbestos boards ½ inch thick, 1 inch high, and 7 inches long. Each coil has a carrying capacity of three amperes and is securely fastened to the bottom in order that the box may be used in any position. Three of these coils are connected to a three-way snap switch A, and three with a similar switch A', Fig. 4. By this means from three to eighteen amperes may be in use at one

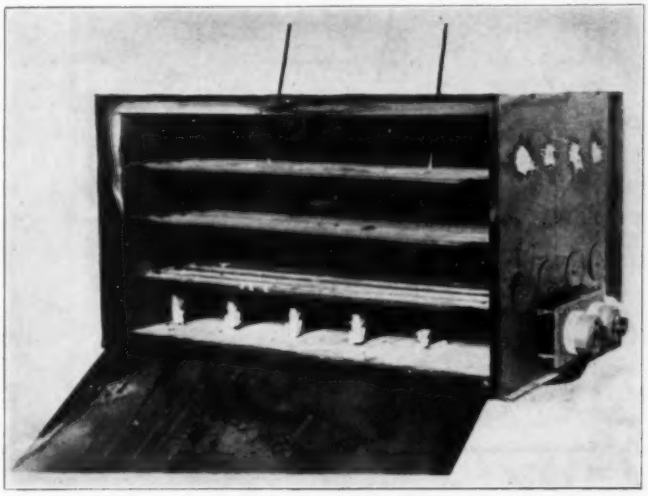

FIG. 1.—A view of the oven

time. The return wire is L, Fig. 4, which is also grounded on the box. The connections of coils Nos. 3 and 5, Fig. 4, pass to the thermo-regulator, T, Fig. 2, at R and R', thence through the screws K and K', Fig. 5, to the platinum contacts Pt, through the strip S to the box and out via the lead wire L to B, the socket, Fig. 4.

The construction of the thermo-regulator is as follows. A strip S, consisting of an above piece of copper 0.022 inch thick, is brazed to a piece of steel of the same thickness, and split longitudinally for an inch back from the free end so as to produce two contacts. S, Fig. 6, is fastened to the under side of block B, Fig. 5, which is pivoted between E and E'. The latter are inset along the

sides of the base plate A. D is a brass rod brazed to the block B, making D–B–S a continuous piece pivoted at V, Fig. 6. S', Fig. 6, is a steel spring pressing firmly against the under side of D so that

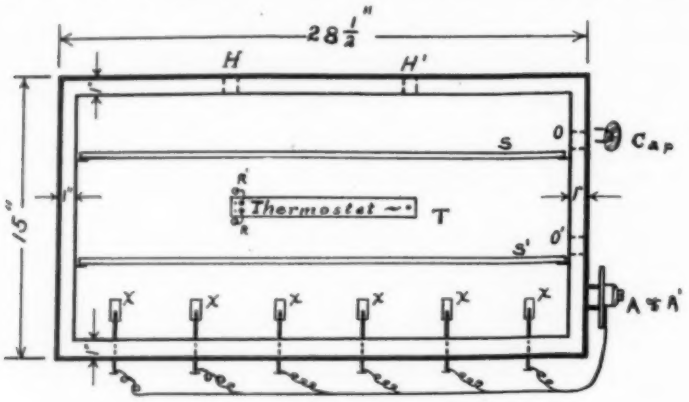

FIG. 2.—Front elevation of the oven

it is held close to the tip of screw C even when the contacts Pt are broken. M and M' are mica strips insulating R and R' from the rest of the device. W is a copper washer. H, Fig. 6, represents holes in the base plate A, through which bolts pass that hold the regulator against the back of the oven. A small condenser, not shown, is shunted around K and Pt, Fig. 6.

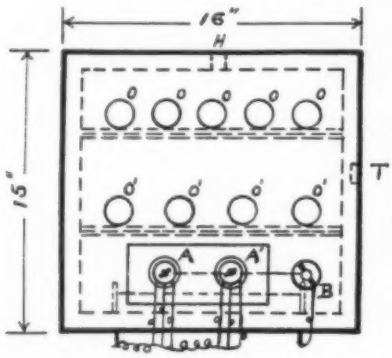

FIG. 3.—End elevation of the oven

Since the temperature coefficient of copper is greater than that of steel, a rise in temperature causes the strip S to bend the contacts Pt away from the tips of the screws K and K', and thus cut out the coils Nos. 3 and 5. The fall of temperature which results causes a reversal of the bend of the strip until the contacts are again made

at *Pt*, and the temperature rises. To adjust the oven for operation at higher or lower temperatures the screw *C*, Fig. 6, is turned in and the oven allowed to come to constant temperature. This requires about one hour. If the temperature so obtained is too low, the screw must be turned in still farther and such adjustments continued until the desired point is reached. Since the sensitiveness of the oven is much greater with both contacts acting together, and the expansion of the metals produces some torsion, it is necessary to adjust the screws at *K* and *K'*, until both coils Nos. 3 and 5

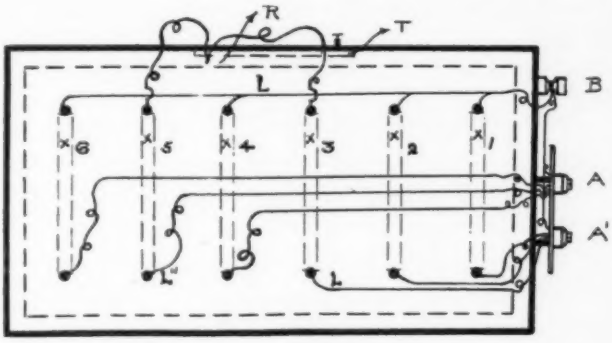

FIG. 4.—Under side of oven, showing wiring and arrangement of coils

come on or off together. Parts of the regulator not otherwise described are made of brass.[1]

This oven has proved highly satisfactory when used on an ordinary 110-volt power or lighting circuit. At the highest temperature at which it has been used, 300° C., the variations during several weeks have not been more than 3°. It has been in use as long as three months continuously, with similar slight variation, at about 180° C. Greater sensitiveness can be obtained by lengthening the strip *S*. The device has also served as an efficient drying oven. When but one of the six coils is being used—preferably No. 4, because it is centrally placed—it maintains a temperature averaging around 110° C., without having the regulator in the circuit.

[1] The writer is indebted to Captain A. De Khotinsky of Kent Chemical Laboratory for the greater part of the work in the design and construction of the oven.

In the early part of the investigation Jena glass tubes were used, provided with an inner tube of copper open at one end, in which the minerals and solutions were placed. This prevented contact of the solutions and the glass when the apparatus was placed in a nearly vertical position, but left only a small part of the solutions in direct contact with the mineral powder. Where glass tubes were used, these were first placed inside of steel tubes made of ordinary steam pipe $1\frac{1}{2}$ inches inside diameter, closed at one end by a screw

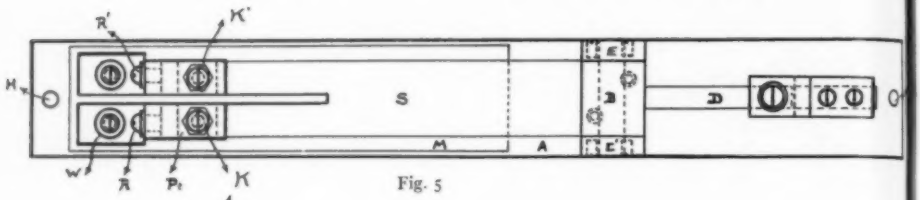

Fig. 5

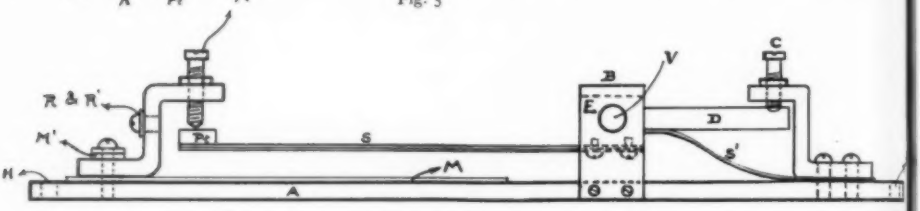

Fig. 6

FIG. 5.—Thermo-regulator, plan. FIG. 6.—Thermo-regulator, side elevation

cap and placed in the oven with the open end against one of the holes o and o', Fig. 3. Accidents to the thermo-regulator from possible explosions were thus prevented, though the trouble caused by bursting of the glass tubes led to their abandonment. In their place copper tubes about 18 to 22 inches long with $\frac{5}{8}$ inch inside diameter and $\frac{1}{16}$-inch walls were used. One end of each of these was closed with a plug of copper about $\frac{1}{8}$ inch thick driven in about $\frac{1}{8}$ inch past the end of the tube and the end then sealed in an oxy-acetylene flame. The solution and the mineral powder were then placed in the tube, a plug of copper inserted as above, and the end then sealed while the tube was partly immersed in a freezing mixture.

These tubes have a calculated bursting strength of 6,000 pounds to the square inch and have proved highly satisfactory. They have been practically unattacked except by sulphide solutions, in which cases the tubes were found lined with beautiful chalcocite crystals.[1] They cost after sealing about one-half as much as Jena bombs of similar size and can be used repeatedly. The oven will accommodate about fifty such tubes at one time.

EXPERIMENTAL WORK

Group I: Effect of pure water on feldspar and hornblende.—In many hydrothermal processes it is not yet clear how significant a rôle is played by the water independent of the dissolved matter. Hence the first step in this investigation was an attempt to determine the efficiency of pure water in such processes. Accordingly the powdered minerals were covered with 300 c.c. of distilled water in open nickel crucibles with a device for maintaining the water at constant volume. The crucibles were placed on electric hot plates

TABLE I

No.	Mineral	Time	Vol. of Solution	Vol. Tube	Temp.	Press. Atm.	Results
1.....	Adularia	14 days	300 c.c.	.....	100° C.	1	No alkalinity
2.....	Hornblende	14	300	.....	100°	1	" "
3.....	Adularia	82	50	80 c.c.	183°	11	" "
4.....	Hornblende	82	50	80	183°	11	" "

and the water kept boiling for 14 days. At the end of that time the solutions were tested for alkalinity with neutral litmus paper and phenolphthalein. No change of color appeared after five minutes' standing. A similar pair of experiments at 180° C. in the sealed copper tubes ran for 82 days and at the end of that time the solutions were similarly tested and again no alkalinity developed. The minerals were then examined under the microscope and compared with slides of untreated mineral (Table I). No change could be detected. These results are not quite in accord with those of

[1] In certain cases a small amount of copper recrystallized on the plugs, probably owing to the fact that the drawn copper has a higher solution tension than crystalline copper.

Clarke[1] and Steiger,[2] who found that the solutions obtained by allowing various minerals—including orthoclase—and rocks to stand in contact with water containing a few drops of phenolphthalein developed more or less alkalinity. Cornu[3] obtained an alkaline reaction toward moist litmus paper from various powdered minerals. Königsberger and Müller[4] found also that adularia was but little attacked at 300° C. by pure water, the chief attack being parallel to the principal cleavage. Up to this temperature water alone is not to be considered an important reagent toward adularia or hornblende.

Group II: Effect of sodium carbonate solutions on feldspars.—The efficiency of carbonate solutions in hydrothermal processes is commonly conceded; this conclusion is based on the abundance of carbon dioxide in certain hot spring waters, and the presence of carbonates in the altered rocks and in the veins. The data given by Clarke[5] and by Peale[6] show not only that carbon dioxide dominates

TABLE II

No.	Mineral	Time	Solution	Concentration	Vol. Sol.	Vol. Tube	Temp.	Press. Atm.	Results
5..	Adularia	3 yrs. 1 mo.	Na_2CO_3	N/2	1,000 c.c.	.....	15° C. *ca.*	1	No change
6..	"	90 days	"	"	300	.....	100° C.	1	" "
7..	"	82	"	"	50	80 c.c.	183°	11	Analcite
8..	"	18	"	"	40	85	233°	30	"
9..	Microcline	18	"	"	40	75	233°	30	"
10..	Albite	18	"	"	40	85	233°	30	Tube burst

in certain thermal waters but that sodium is generally an abundant metallic ion in such waters. The concentration is low, rarely reaching as much as 1 per cent, though the carbonates make up as much as or even more than three-fourths of the total dissolved matter. In order then to follow the original plan and to use solutions comparable to those found in nature, solutions of sodium carbonate were employed as shown in Table II. The concentration

[1] F. W. Clarke, *Jour. Am. Chem. Soc.*, XX, 739–42.

[2] George Steiger, *ibid.*, XXI, 437–39.

[3] F. Cornu, *Tschermak's Mitteilungen*, XXIV, 417–32. [4] *Op. cit.*, p. 360.

[5] F. W. Clarke, *Bull. 491, U.S. Geol. Survey*, pp. 180–81.

[6] A. C. Peale, *Bull. 32, U.S. Geol. Survey.*

of one-half normal is equivalent to nearly 2.8 per cent. The minerals in Nos. 5 and 6 showed no change under the microscope. Tube No. 10 burst owing to defective sealing, but the mineral was unchanged. In all the others crystals of analcite formed along the sides of the tubes and around grains of feldspar as nuclei. Solution No. 6 was not saved, but the solutions from Nos. 7, 8, and 9 were placed in paraffin-lined bottles and on examination after several weeks' standing showed heavy gelatinous precipitates. These precipitates were then examined and were found to consist of silica and water. In spite of the fact that tube No. 10 burst, the results obtained from later experiments, notably Nos. 33 and 34, make it certain that similar results would have been secured had no accident occurred.

These facts indicate definite attack and solution of the feldspars with loss of one molecule of silica from each molecule of feldspar together with an exchange of the potash of the adularia for the soda of the solution. The following equations, using empirical formulae, indicate the reactions, which have positive experimental basis.

$$KAlSi_3O_8+H_2O+Na' \rightarrow NaAlSi_2O_6,H_2O+SiO_2+K'$$

and

$$NaAlSi_3O_8+H_2O+Na' \rightarrow NaAlSi_2O_6,H_2O+SiO_2+Na'.$$

The analcite was identified optically and also by sifting out the unchanged feldspar with its adhering analcite, gelatinizing the remaining crystals with hydrochloric acid, and allowing a small portion of the solution to crystallize out under the microscope as sodium chloride; the remaining portion of the solution was tested for alumina with ammonia water. After drying at 110° C., the crystals yielded water in a closed tube.

Group III: Sodium carbonate solutions on hornblende.—A series of experiments, Nos. 11, 12, and 13, exactly like those in Group II, were then tried with hornblende as the mineral (Table III). In no

TABLE III

No.	Mineral	Time	Solution	Concentration	Vol. Sol.	Vol. Tube	Temp.	Press. Atm.	Results
11..	Hornblende	3 yrs. 1 mo.	Na_2CO_3	N/2	1,200 c.c.	.....	15° C. *ca.*	1	No change
12..	"	90 days	"	"	300	.....	100° C.	1	" "
13..	"	82	"	"	60	85 c.c.	183°	11	" "

case did there appear, under the microscope, to be any change. No gelatinous silica separated from the solutions after long standing.

Group IV: Potassium fluoride solutions on feldspars and hornblende.—Fluorine has long been looked upon as one of nature's important mineralizers;[1] it occurs in small quantities in the emanations from Kilauea,[2] is abundant as fluorite in metalliferous veins, is a constituent of many minerals such as apatite, amblygonite, lepidolite, topaz, and cryolite, and is considered by Spurr[3] an essential constituent of muscovite. Table IV shows the experiments con-

TABLE IV

No.	Mineral	Time	Solution	Concentration	Vol. Sol.	Vol. Tube	Temp.	Press. Atm.	Results
14..	Adularia	82 days	KF*	N/10	60 c.c.	85 c.c.	183° C.	11	No change
15..	"	18	"	"	50	80	233°	30	" "
16..	Microcline	18	"	"	40	75	233°	30	Minute rods
17..	Albite	18	"	"	30	70	233°	30	No change
18..	Hornblende	82	"	"	60	85	183°	11	Brown iron oxide
19..	"	15	"	"	50	75	233°	30	Tube burst

*This salt contained a small amount of the acid salt HKF_2.

ducted with potassium fluoride. In experiments Nos. 14, 15, 16, and 17 the feldspars showed no change other than a possible slight etching. In No. 16 there appeared some minute bacteria-like rods that were not further identifiable. In No. 18 the hornblende was vigorously attacked. The product is an amorphous brown mass resembling limonite and consisting of hydrated iron oxide together with grains of partially altered mineral. Some of these grains are bleached to isotropic transparency, others have a rim of isotropic matter surrounding ellipsoidal grains in the interior. No gelatinous precipitate appeared in the decanted solution on standing in paraffin lined bottles for several months. In No. 19 the tube burst and the mineral appeared unchanged.

Group V: Mixtures of carbonate and fluoride solutions on feldspars and hornblende.—Since the feldspars were not visibly attacked by the fluoride solutions but had been attacked by the alkaline carbo-

[1] See especially C. Doelter, *Allgem. chem. Min.*, p. 207, and W. Bruhns, *Neues Jahrb. Min. Geol.*, II (1889), p. 26.

[2] A. L. Day and E. S. Shepherd, *Bull. Geol. Soc. Am.*, XXIV, 592.

[3] J. E. Spurr, *Professional Paper 42, U.S. Geol. Survey*, p. 233.

nate solutions, it was deemed advisable to try a mixture of the two solutions to see if the traces of fluoride present would modify the results in any way (Table V). The feldspars in Nos. 20, 21, and 22

TABLE V

No.	Mineral	Time	Solution	Concentration	Vol. Sol.	Vol. Tube	Temp.	Press. Atm.	Results
20..	Adularia	90 days	Na_2CO_3+KF	N/2+N/10	300c.c.	.....	100° C.	1	No change
21..	"	41	" "	" "	54	80c.c.	183°	11	" "
22..	"	82	" "	" "	60	80	183°	11	" "
23..	"	18	" "	" "	40	85	233°	30	Analcite
24..	"	15	" "	" "	60	100	280°	65	Needles
25..	Orthoclase	15	" "	" "	50	100	280°	65	Twins of unknown mineral, also needles
26..	Microcline	18	" "	" "	48	100	233°	30	No positive change
27..	Albite	18	" "	" "	48	80	233°	30	Analcite
28..	Hornblende	90	" "	" "	300	.....	100°	1	No change
29..	"	41	" "	" "	60	85	183°	11	" "
30..	"	82	" "	" "	60	85	183°	11	" "
31..	"	3	" "	" "	60	85	233°	30	" "

were not visibly attacked and the decanted solutions gave no precipitate on standing. No. 23 showed crystals of analcite. No. 24 showed some needles with parallel, or nearly parallel, extinction and possibly some isotropic forms, though these were not positively identified. No. 25 contained many needles like those in No. 24 with no analcite. These needles have an extinction angle of less than 2°; index in the direction of elongation is 1.490, and at right angles to this 1.517; elongation is negative. The crystals after drying at 110° C. yield no water in a closed tube. Good terminations at both ends are common. No. 25 also contained beautifully twinned crystals, illustrated in the accompanying sketch (Fig. 7) made with a camera lucida. The crystals

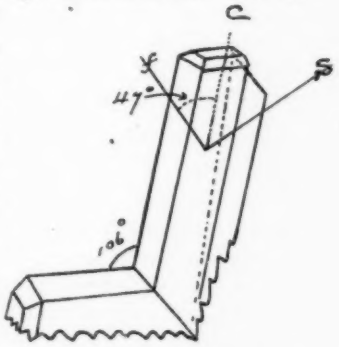

FIG. 7.—Twinned crystals produced by experiment No. 25.

are perfectly transparent, have an index close to but greater than 1.565, an extinction angle of 43°, and are probably monoclinic. In experiment No. 26 no evidence of any change appeared. In experiment No. 27 icositetrahedra of analcite appeared, some free and some including fragments of the feldspar. The hornblende in all of this group of experiments was unchanged. It is concluded that the presence of the fluoride had practically no influence upon the alteration.

Group VI: Sodium bicarbonate solution upon feldspars and hornblende.—These experiments embody an attempt to increase the pressure both by raising the temperature and by increasing the concentration of the carbon dioxide through dissociation of bicarbonates (Table VI). In experiment No. 32 scarcely any feldspar

TABLE VI

No.	Mineral	Time	Solution	Concentration	Vol. Sol.	Vol. Tube	Temp.	Press. Atm.	Results
32..	Adularia	15 days	$NaHCO_3$	N/2	45 c.c.	85 c.c.	233° C.	30	Analcite
33..	"	15	"	"	50	85	280°	65	"
34..	Albite	15	"	"	45	85	233°	30	Analcite and needles
35..	"	15	"	"	60	105	280°	65	Tube burst, but needles formed
36..	Hornblende	15	"	"	45	85	233°	30	No change
37..	"	15	"	"	45	85	280°	65	" "

remained and much analcite appeared as free crystals and as aggregates. Gelatinous silica appeared in the decanted solutions after some time. Similar results with decided etching of the feldspar grains appeared in No. 33 where the pressure was practically doubled. Possibly less analcite formed in this than in the previous experiment at the lower temperature. In No. 34 with albite, analcite also formed with many needles like those in experiment No. 25, Table V. In experiment No. 35 the tube burst where poorly sealed, but the mineral was nearly all altered to needles as in experiments Nos. 24, 25 and 34. These were analyzed qualitatively and found to consist of soda, alumina, and silica and yielded no water in a closed tube. The writer was not able to find any natural sodium aluminum silicate whose properties agree with these. Needles having

the same optical properties as these also appeared in experiment No. 46 and are undoubtedly the same thing.

The hornblende was unchanged in experiments No. 36 and 37; the tube in No. 37 burst some time during the course of the heating.

Group VII: Potassium bicarbonate solutions on feldspars and hornblende.—These are similar to the experiments of Group VI, except that potassium bicarbonate solutions were substituted for the sodium bicarbonate solutions (Table VII). In no case either

TABLE VII

No.	Mineral	Time	Solution	Concentration	Vol. Sol.	Vol. Tube	Temp.	Press. Atm.	Results
38..	Adularia	15 days	$KHCO_3$	N/2	45c.c.	75c.c.	233° C.	30	No change
39..	"	15	"	"	50	90	280°	65	" "
40..	Albite	15	"	"	45	75	233°	30	" "
41..	"	15	"	"	60	105	280°	65	" "
42..	Hornblende	15	"	"	45	100	233°	30	" "

with the soda or potash feldspar or with hornblende did any change appear in the minerals. The writer is unable to give an adequate explanation of this fact, but it may have some bearing on the question whether potash is introduced or not in hydrothermal processes.

Group VIII: Sodium tetraborate solution upon feldspars and hornblende.—A few experiments were tried with borax solutions with results very similar to those produced by the alkali carbonates (Table VIII). In No. 43 well-formed crystals of analcite as rhom-

TABLE VIII

No.	Mineral	Time	Solution	Concentration	Vol. Sol.	Vol. Tube	Temp.	Press. Atm.	Results
43..	Adularia	15 days	$Na_2B_4O_7$	N/4	50c.c.	100c.c.	233° C.	30	Analcite
44..	Albite	15	"	"	50	95	233°	30	No change
45..	Hornblende	15	"	"	50	90	233°	30	" "

bic dodecahedra appeared. No alterations occurred in the other experiments.

Group IX: Sodium sulphide solutions upon feldspars and hornblende.—The presence of the metallic sulphides indicates that at

certain phases of the vein-forming process sulphide solutions must be present. Though the solutions are undoubtedly very complex, their efficiency as hydrothermal agents is very probably due to a few components. The physical state of these solutions has been recently shown by Tolman and Clark[1] to depend decidedly upon the composition, at least at ordinary temperatures, and it may also have some decided influence upon the character of the alteration. In the following experiments the copper tubes were vigorously attacked and chalcocite crystals lined the walls of the tubes (Table IX). In No. 46 needles of the anisotropic crystals like those

TABLE IX

No.	Mineral	Time	Solution	Concentration	Vol. Sol.	Vol. Tube	Temp.	Press. Atm.	Results
46..	Adularia	15 days	Na_2S	N/2	50c.c.	120c.c.	233° C.	30	Analcite and needles
47..	Albite	15	"	"	50	100	233°	30	Analcite
48..	Hornblende	15	"	"	50	105	233°	30	No change

obtained in experiments Nos. 25 and 34 were formed, together with well-formed analcite crystals. Very little of the original feldspar remained. In No. 47 perfect analcite crystals appeared as icositetrahedra and as combinations of the cube and rhombic dodecahedron. These vary in size from one-half to one millimeter in diameter. The hornblende in experiment No. 48 was not attacked and no pyrite could be identified in the product. No sulphur was obtained by heating the mass in a closed tube. The results with the feldspars are quantitatively greater than in any other experiment; this was probably due to the fact that the hydrogen sulphide is a weaker acid than is carbonic acid and the hydrolysis therefore produces a more strongly alkaline solution.

Group X: Aluminate solutions on feldspars and hornblende.—In the previous experiments the loss of silica from the minerals resulted in an apparent rise of the alumina content of the new minerals. With the thought that possibly an increase in the concentration of the alumina in the solutions might cause the solubility product for

[1] C. F. Tolman and J. D. Clark, *Econ. Geol.*, IX (1914), 559.

compounds richer in alumina to be exceeded, a few experiments with aluminate solutions were tried. These solutions were prepared by taking a weighed quantity of aluminum sulphate, precipitating the aluminum as hydroxide, washing the precipitate and then adding it to normal sodium or potassium hydroxide and diluting

TABLE X

No.	Mineral	Time	Solution	Concentration of K or Ka	Vol. Sol.	Vol. Tube	Temp.	Press. Atm.	Results
49..	Adularia	15 days	Sodium aluminate	N/2	40c.c.	80c.c.	280° C.	65	Analcite
50..	"	15	Potassium aluminate	"	40	85	280°	65	Hexagonal plates and needles
51..	Orthoclase	15	Potassium aluminate	"	60	120	280°	65	Hexagonal plates and needles
52..	Albite	15	Potassium aluminate	"	30	70	280°	65	Hexagonal plates
53..	Hornblende	15	Potassium aluminate	"	35	75	280°	65	No change

one-half. After this had been allowed to stand for several hours it was filtered from its slight precipitate (Table X). In No. 49 analcite crystals formed, which were identified chemically and microscopically. In Nos. 50, 51, and 52 twinned hexagonal plates, with anomalous division into fields under polarized light, resulted, as shown in Fig. 8. These gelatinize with hydrochloric acid and contain sodium, but no aluminum could be detected in them by a microchemical test. The hornblende was not attacked.

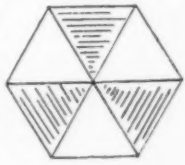

FIG. 8.—Twinned hexagonal plate resulting from experiments Nos. 50, 51, and 52.

Group XI.—Albite and hornblende were heated for 15 days at 280° C., with saturated solutions of calcium bicarbonate. The tubes burst in each case and the minerals showed no change.

GEOLOGIC BEARING

Though the alteration of feldspars to analcite has not been commonly described, in many cases it is very possible that some of the

determinations of isotropic minerals as glass may be incorrect, and that analcite has been overlooked. In examining the slides in the University of Chicago collection a slide from a trachyte of Bannberscheid, Westerwald, Nassau, was found which showed the soda feldspars altered to analcite though the original crystal boundaries remained sharp. The alteration does not follow cleavage cracks but appears in irregular patches. Mr. K. F. Mather in a forthcoming paper will describe an eruptive cone of Quaternary age in the canyon of the Mancos River ten miles southwest of Mancos, Colorado—locally known as the "blowout"—which is cut by dikes of augite minette. These dikes carry fragments of granite xenoliths which are deeply corroded and partially assimilated. The feldspars are altered to analcite—identified microchemically. A careful study of rock specimens would probably show that this type of alteration is much more common than has been supposed.

SUMMARY

1. Alkaline solutions of different characters dissolve the feldspars with separation of silica and crystallization of compounds less rich in silica. The solutions are probably hydrolyzed since the reactions are accelerated by the presence of alkalies, by increased concentration of the alkalies, and by higher temperatures.

2. Feldspars and hornblende are not appreciably attacked by pure water at temperatures up to about 300° C. adularia to at least 350° C., showing that the dissolved substances rather than water alone must cause the differences in the nature of the alterations.

3. Albite and orthoclase feldspar seem to respond to the action of the alkalies in nearly identical ways, and hence the conclusion is patent that they have very similar chemical structures.

4. The influence of small amounts of fluoride and borates as mineralyzers has not been found important, at least in the presence of the other substances. This leads to the suggestion that possibly the mineralyzing effect is merely that of causing solution at temperatures where the silicates in question would otherwise be much less soluble.

5. It is notable that no kaolin or kaolin-like substance forms from alkaline solutions at temperatures up to 280° C. The sugges-

tion seems necessary that since pure water has practically no effect on the feldspars, and that since the alkaline waters produce minerals other than kaolin, kaolin probably forms by the action of acid solutions upon the feldspars. The literature bearing on the field occurrences of kaolin shows a striking lack of references to association of carbonates and kaolin, though this would be expected if carbonated waters are the cause of the formation of kaolin from the feldspars.

6. The general agreement of the data obtained throughout the range of temperatures used shows that the silicates may be studied with the apparatus described, up to 300° C., with gratifying results and without great mechanical difficulties, and also without the necessity of contamination from undesirable sources such as glass tubes.

The writer is indebted to Professor W. H. Emmons for suggesting the problem and to Professor A. D. Brokaw for sincere interest and many suggestions during the progress of the work. Further work of a similar sort is in progress and the next paper will deal with the action of various acid solutions, especially hydrofluoric acid on the same group of minerals.

ZONAL WEATHERING OF A HORNBLENDE GABBRO

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In connection with a study of the alteration of the so-called trap or diabase dikes at La Grange, Georgia,[1] one of the writers collected some interesting specimens showing a weathered zone in which the decomposition is extreme, with an abrupt transition into fresh rock, practically free from the effects of weathering. This paper is concerned with the description of a typical specimen, and with the results of chemical analyses of the fresh rock, the partly altered material, and the extreme phase of alteration present.

La Grange is in the extreme western edge of the state, not many miles from the southern end of the belt of crystalline rocks which extends from Maine to Alabama. The country rock is for the most part gneiss, cut by granitic intrusions. Both gneiss and granite are cut by pegmatites and basic dikes, the latter usually called diabase, and supposed to be of Triassic age. At La Grange there is a series of dikes, only a few feet apart, ranging up to forty feet in width, with an average of about four to six feet. The dikes have a general north-south trend, and are abundant over an area of a little more than half a square mile.

The older crystalline rocks are very deeply weathered. The basic dikes have been considerably altered at the surface, and in many instances may be traced by the strong iron stain they have imparted to the soil. In some excavations, however, material showing no appreciable effects of weathering may be obtained. In some cases weathering has been strongly marked along joints, changing the character of the material for an inch or less, beyond which the rock is fresh. Fig. 1 shows a specimen in which the zone of alteration is about an inch thick. Analyses were made of the

[1] L. P. Smith, *Alteration of Diorite by Weathering*, Dissertation, University of Chicago, 1915.

outside, much-altered portion, and of the part of the weathered zone adjacent to the fresh material. A specimen from an entirely fresh portion of the same dike was analyzed for a comparative study.

The fresh rock is finely grained, holocrystalline, nearly black in color, with small white feldspar crystals evenly distributed through

FIG. 1.—Specimen showing zone of weathering $\times\frac{1}{2}$. Circle shows approximately the part from which the slide shown in Fig. 3 was taken.

the mass, which is dominantly hornblende. Small garnets, irregularly distributed, and small crystals of pyrite are present. Microscopically, the rock is found to consist of hornblende (65 per cent), which may be secondary, labradorite (32 per cent), with small amounts of orthoclase and augite, and accessory magnetite, pyrite, titanite, and apatite. One of the typical slides is shown in Fig. 2. The rock is a hornblende gabbro.

The partly altered material is notably lighter in color than the fresh rock. Most of the hornblende has been weathered, leaving brown iron stains and a small amount of chlorite, but some particles of apparently fresh hornblende are to be seen. The feldspar has been changed to a chalky white material in an iron-stained matrix. Microscopical study reveals some unchanged hornblende, but for the most part only alteration products are discernible. Limonite, white mica (probably also gibbsite), chlorite, zoisite, and a small quantity of magnetite make up the mass. The weathered zone is sharply set off from the fresh material even in thin section. In Fig. 3 the fresh portion is shown on the left, changing to altered on the right.

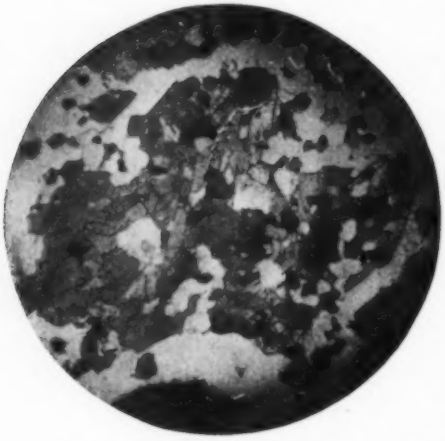

FIG. 2.—Section of fresh hornblende gabbro ×20. (Two rather large cracks appear as light bands.) Ordinary light.

The most completely weathered portion is a friable, earthy, non-plastic mass, strongly iron stained, and so thoroughly disintegrated that the original texture of the

FIG. 3.—Slide showing transition from fresh to altered portion of specimen. ×25. Crossed nicols.

rock is practically lost. No microscopical study of this material was undertaken.

The results of chemical analyses, made in duplicate in every case and in triplicate in a number of determinations, are given in Table I. The following recognized effects of weathering are well illustrated: loss of silica, apparent increase in alumina, increase in ferric iron, decrease in ferrous iron, loss of bases, increase of combined water, and decrease in specific gravity. The very small development of carbonates may be due to the fact that most of the available carbon dioxide has formed soluble bicarbonates in removing lime, magnesia, and alkalies.[1]

TABLE I

	1*	2†	3‡		1*	2†	3‡
SiO_2	45.16	26.02	23.34	TiO_2	.31	.18	.15
Al_2O_3	17.52	28.60	32.70	MnO	.47	.51	tr.
Fe_2O_3	3.12	11.38	21.77	CO_2	none	.12	.52
FeO	6.99	4.00	none	S	.10	.04	none
MgO	4.67	3.03	.57				
CaO	17.50	7.96	.75	Total	100.10	100.23	100.20
Na_2O	2.39	1.54	.39				
K_2O	1.37	.81	1.19	Total Fe	7.62	11.78	15.24
H_2O-	.04	.87	3.77	Sp. Gr.§	3.020	2.813	2.340
H_2O+	.46	15.17	15.05				

* Fresh Rock.
† Altered near the fresh rock.
‡ Altered, most decomposed portion.
§ Pycnometer method.

On the assumption of constancy of alumina[2] the analyses may be recalculated, and gains and losses estimated, as shown in Table II. The change in total iron is worthy of note, in that there is a slight loss shown in 2, with a marked increase in 3. A suggested explanation is that during the early stages of alteration some soluble ferrous compound was formed, which migrated, perhaps by capillary action, to the outer zone before it was precipitated by oxidation to

[1] In this connection it may be noted that wells in or near the dikes are said to yield water containing considerable amounts of lime, while those in the gneiss, farther from the dikes, yield comparatively soft water. Unfortunately, no quantitative data on these waters are available.

[2] G. P. Merrill, *Rocks, Rock Weathering and Soils*, p. 208.

the ferric condition. That the removal of ferrous iron may take place in the early stages of alteration has been commonly recognized.[1]

TABLE II

	1	2	Gain	Loss	3	Gain	Loss
SiO_2	45.16	15.94	.....	29.22	12.51	.....	32.65
Al_2O_3	17.52	17.52	.....	.....	17.52	.....	.....
Fe_2O_3	3.12	6.97	3.85	.....	11.66	8.54	.....
FeO	6.99	2.45	.....	4.54	none	.....	6.99
MgO	4.67	1.86	.....	2.81	.31	.....	4.36
CaO	17.50	4.87	.....	12.63	.34	.....	17.16
Na_2O	2.39	.94	.....	1.45	.21	.....	2.18
K_2O	1.37	.50	.....	.87	.64	.....	.73
H_2O-	.04	.53	.49	.....	2.36	2.32	.....
H_2O+	.46	9.29	8.83	.....	8.06	7.60	.....
TiO_2	.31	.11	.....	.20	.08	.....	.23
MnO	.47	.31	.....	.16	tr.	.....	.47
CO_2	none	.07	.07	.....	.28	.28	.....
S	.10	.02	.....	.08	none	.....	.10
Total	100.10	61.39	.....	.....	53.95	.....	.....
Total Fe	7.62	7.21	.....	4.05	9.33	1.71	.....

Essentially the same relations are shown by the "straight line" diagrams of Mead.[2] In Fig. 4 the analyses of the altered portion are compared with that of the fresh rock. The full line represents analysis 2, and the broken line analysis 3. In general, 3 is merely an accentuation of 2, except for potash and total iron, both of which show a change of sign in the direction of change. It is apparent that in the later stage of alteration the removal of bases continues, and that it is out of proportion to the further removal of silica as compared with 2. The retention of potash is by no means unusual.

In Table III the analyses are recast to molecular proportions to emphasize some of the chemical and mineralogical features. It is to be noted that even in analysis 2 the amount of silica is insufficient to combine with all of the alumina to form kaolin. The alumina : silica ratios are as follows: 1, 1:4.36; 2, 1:1.57; 3, 1:1.19. For kaolin 1:2 is required. On the extreme assumption that all of the silica in 3 is present in the form of kaolin, the analysis may be said to

[1] C. K. Leith and W. J. Mead, *Metamorphic Geology*, p. 22, Henry Holt & Co. (1915).

[2] W. J. Mead, *Econ. Geol.*, VII (1912), 141–44.

represent (by weight) kaolin, 50 per cent; limonite, 25 per cent; bauxite, 18 per cent; other substances, 7 per cent. It is highly probable that part of the silica is present in some compound not containing alumina, and if so the amount of bauxite is correspondingly greater. This lends a distinctly bauxitic-lateritic aspect to the alteration.

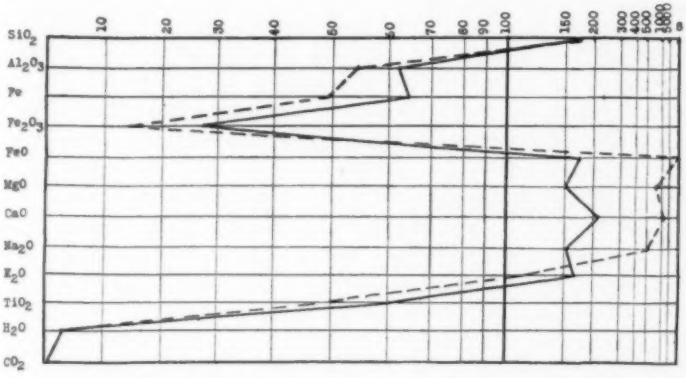

FIG. 4.—"Straight line diagram" showing changes in composition with alteration. The full line represents analysis 2, the broken line analysis 3, compared with the analysis of the fresh rock.

TABLE III

Per cent ÷ mol. wt.

	1	2	3		1	2	3
SiO_2	.747	.431	.387	CaO	.312	.142	.013
Al_2O_3	.172	.280	.320	Na_2O	.039	.025	.006
Fe_2O_3	.020	.070	.136	K_2O	.015	.008	.013
FeO	.097	.056	.000	H_2O+	.026	.845	.836
MgO	.116	.075	.014	CO_2	.000	.003	.012

Specimens similar to the one studied are not uncommon in the weathered part of the dikes, and the changes shown are believed to be fairly typical for the rock in question. It is to be noted that the fresh rock is much richer in iron than rocks which yield bauxite deposits, and richer in alumina than those yielding high grade laterite deposits.

Reviews

Lockatong Formation of the Triassic of New Jersey and Pennsylvania. By A. C. Hawkins. Annals N.Y. Acad. Sci., XXIII. 145–76, Plate 1, January 27, 1914.

The Lockatong formation is the middle member of the Newark series of the Triassic, extending from a point just west of Phoenixville, Pennsylvania, to Princeton, New Jersey. The rocks of the formation are dense, fine-grained, massive argillites, with some shales. The formation as a whole has a decidedly lens-like character. On the basis of the general structure, lithologic character, and type of fossils, which include estheriae, fish-scales, ostracods, and plant remains, it is concluded that the sediments were laid down near the center of an inland basin. The particles of the argillite are for the most part cemented by silica, which renders the formation very hard and a pronounced ridge-maker. The color of the beds is due to iron in various states of oxidation. The boundaries of the Lockatong are very uncertain, owing to the fact that it passes by a series of transitional dovetailing strata into the other formations of the Newark. Since part or all of the Lockatong may be contemporaneous with portions of the Stockton and Brunswick formations elsewhere, it seems that as a definite geological time unit the Lockatong is valueless. There are three principal joint directions in the Lockatong formation, the most important of which is remarkably constant, and extends into the borders of a diabase mass near Rocky Hill, which is interpreted as an extension of the Palisade sill. Titanium minerals, brookite and ilmenite, are found in this major joint series, apparently far removed from the diabase. Analcite and barite also occur. That these minerals are derived from the igneous rocks is indicated by similar occurrences in New Jersey elsewhere. Parts of the Lockatong argillite are very well adapted for commercial use.

R. C. M.

Geological Map of Tennessee. Compiled by Olaf P. Jenkins, A. H. Purdue, State Geologist.

This map represents Archean, Cambrian, Ordovician, Silurian, Devonian, Mississippian, Pennsylvanian, Cretaceous (Upper), Eocene, Pleistocene, and Recent formations. Few states have so wide a range

of systems, Proterozoic, Permian, Lower Cretaceous, Miocene, and Oligocene only being absent. Pliocene (Lafayette) is indicated in the legend of the map but not shown on the map itself, and the legend seems to be intended to throw doubt on the validity of the formation in Tennessee.

Under the designation "Columbia Formation," loess, loam, and loose sand are grouped. This seems to us an unfortunate classification. The "Terrace Deposits" of the map are quite as appropriately classed as "Columbia" as the loess and loam which are so classed. We are of the opinion that the use of the term "Columbia Formation" should be discontinued (though possibly the term "Columbia Series" may be useful to include all Pleistocene non-glacial formations). What was originally grouped under the name Columbia included several formations of which the probable equivalents of the Terrace Deposits of this map were a chief member. "Loess" would seem to be an adequate designation of the deposits included under that term, without classing them as Columbia. Their classification as Pleistocene seems altogether adequate. The loess, of many regions at least, is of very different ages, and all of it does not belong to one formation in the chronological sense.

The map is distinct and represents sufficient change from its predecessors to be welcome. It is accompanied by elaborate explanatory legends and by four cross-sections which represent well the structure of the formations in the state.

The map may be had by application to the State Geologist, Nashville, Tennessee. Postage, 8 cents.

R. D. S.

Cretaceous Deposits of the Eastern Gulf Region, and Species of Exogyra from the Eastern Gulf Region and the Carolinas. By L. W. STEPHENSON. U.S. Geol. Surv., Prof. Paper 81, 1914. Pp. 75, pls. 21, charts 8.

In eastern Alabama and Georgia a terrane, previously regarded as forming the eastward extension of the Tuscaloosa series of western Alabama, has been shown by its unconformable relations with overlying formations, lithologic character, and contained plant fossils to be of Lower Cretaceous (Comanchean) age, though probably somewhat younger than the Patuxent. Belonging to the Upper Cretaceous (Cretaceous) of the eastern Gulf region are four formations, Tuscaloosa (regarded as Lower Cretaceous), Eutaw, Selma chalk, and

Ripley. The first, consisting of irregularly bedded sands, clays, and gravels, has an estimated thickness of 1,000 feet and rests unconformably on a basement of Paleozoic metamorphics, and in the east on Pre-Cambrian crystalline rocks and in part on Lower Cretaceous. The Eutaw formation, somewhat similar to the Tuscaloosa in lithologic character, is believed to be entirely marine, though much of the formation was doubtless laid down in very shallow water. It is 400–500 feet thick, rests conformably on the Tuscaloosa, and is overlain conformably in part by the Selma chalk, and in part by the Ripley formation. The Selma chalk consists mainly of more or less argillaceous and sandy limestones rendered chalky by their large content of foraminiferal remains. It is abundantly fossiliferous in certain portions, yielding large numbers of the *Exogyra* described in the latter portion of the paper. The Selma grades into the sandy member of the Tuscaloosa, and the clastic beds of the Ripley formation when followed along the strike. A thickness of 930 feet of the chalk formation has been measured in western Alabama. The Ripley formation, 250–350 feet in thickness, consists typically of calcareous and glauconitic sands, sandy clays, and impure limestones and marls of marine origin. It extends through parts of the Gulf states from southern Illinois to Georgia. A study of the faunas of the various formations is detailed, and correlations with other Cretaceous regions indicated by chart.

A description of the genus *Exogyra*, which includes three species with two varieties, constitutes the second portion of the paper.

R. C. M.

The Jurassic Flora of Cape Lisburne, Alaska. By F. H. KNOWLTON. U.S. Geol. Surv., Prof. Paper 85, Part D, 1914. Pp. 25, pls. 4.

The Jurassic of the Cape Lisburne area is estimated to have a very great thickness, 15,000 feet, and contains from 40 to 50 coal beds which range in thickness from 1 or 2 feet to over 30 feet. Plant collections from this area show 17 species of well-defined Jurassic types. The close similarity or identity of a number of forms with species from eastern Siberia and Mongolia is noteworthy. The flora indicates a warm-temperate or subtropic climate and the geographic range, especially into the Arctic and Antartic, is suggestive of the uniform mildness of the Jurassic earth-climate.

R. C. M.